Volume 25 SEPTEMBER, 1955 Number 3

Journal of Sedimentary Petrology

A Publication of the Society of Economic Paleontologists and Mineralogists
a Division of
The American Association of Petroleum Geologists

« CONTENTS »

Mineralogy and petrology of the rocks near the Quadrant-Phosphoria boundary in southwest Montana ..... *Charles E. Weaver* 163

Clay minerals of the Neuse River estuary ..... *George M. Griffin and Roy L. Ingram* 194

Sandstone classification: relation to composition and texture ..... *John Bokman* 201

The nature of corrosion zones in the Middle Ordovician of Minnesota ..... *N. Prokopovich* 207

Diagnostic heavy minerals in the regolith of Massachusetts ..... *Mitchell A. Light* 216

Differential surface weathering of Bighorn dolomite ..... *Clayton H. Johnson and Donald L. Biggs* 222

Extraordinarily coarse eolian sand of the Ica Desert, Peru ..... *Norman D. Newell and Donald W. Boyd* 226

Summary of "Concerning some additional aids in studying sedimentary formations" by M. S. Shvetsov ..... *Richard D. Terry and George V. Chilingar* 229

Hollow sanidine grains: stratigraphic marker for the Pierce Canyon formation, west Texas and southeastern New Mexico ..... *D. N. Miller, Jr.* 235

ANNOUNCEMENTS

Advisory committee on radioactive mineral exploration ... 238

General bathymetric chart of the oceans 239

Permian Basin Section field trip 240

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The nature of corrosion zones in the Middle Ordovician of Minnesota ..... *N. Prokopovich* 207

Diagnostic heavy minerals in the regolith of Massachusetts ..... *Mitchell A. Light* 216

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Permian Basin Section field trip 240

NOTICE

Articles submitted for publication to the *Journal of Sedimentary Petrology* should have a prefatory abstract and sufficient postage should be sent by the author to pay the return of the manuscript as it has been found that many articles require more or less revision before publication.

Persons who are not members of the Society of Economic Paleontologists and Mineralogists or are not subscribers to the *Journal of Sedimentary Petrology* but who desire to publish an article in the *Journal,* should have some member of the Society of Economic Paleontologists and Mineralogists or some member of the American Association of Petroleum Geologists sponsor the article. Members of these Societies are given preference in publication.

Among papers to appear in forthcoming issues of the *Journal* are the following:

Petrography of the late Tertiary volcanic ash falls in the central Great Plains. By Ada Swineford, John C. Frye, and A. Byron Leonard.

A theoretical curve for statistical analysis of sediments. By E. Alan Lohse.

Fucoidal markings in the Swan Peak formation, southeastern Idaho. By Henry W. Coulter.

Phi-millimeter conversion table. By Harry G. Page.

Inverse graded bedding in primary gypsum of chemical deposition. By L. Ogniben.

Student operator error in determination of roundness, sphericity, and grain size. By Robert L. Folk.

Mounting technique for grain size and shape measurement. By C. W. H. Hulbe.

JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 25, NO. 3, PP. 163–192
FIGS. 1–11, SEPTEMBER, 1955

MINERALOGY AND PETROLOGY OF THE ROCKS NEAR THE QUADRANT-PHOSPHORIA BOUNDARY IN SOUTHWEST MONTANA

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ABSTRACT

A petrographic study was made of the rocks near the boundary of the lower Pennsylvanian Quadrant and the Permian Phosphoria formations. The mineralogy is described in detail and some of the detrital minerals, quartz, chert, tourmaline, and zircon, are subdivided into a variety of species. The chemical minerals described in detail are chert, calcite, dolomite, collophane, kaolinite, and illite.

The Quadrant in this area is composed largely of fine-grained siltstones (dolomitic, illitic, and kaolinitic) and carbonate rocks. There is little significant difference in the detrital minerals of the two formations. However, the Phosphoria contains a much larger percentage of chemical rocks than the Quadrant.

The detritus of the two formations was primarily derived by the reworking of older sediments. The minerals of the two formations, in this basin-edge area, comprise a quartzite-limestone series and are typical of a period of near peneplaination in the source area and belong in the shallow water marine facies of sedimentation. It is indicative of epeirogenic tectonic control both in the source area and in the locus of deposition. The rocks in the Phosphoria are modified by the deposition of an abundance of authigenic minerals and a decrease in detrital minerals. This modification was likely caused by an abundance of volcanic material in the source area.

INTRODUCTION AND GENERAL GEOLOGY

Statement of the Problem

In southwest Montana the boundary between the Quadrant formation of lower Pennsylvanian Age and the Phosphoria formation of Permian Age is difficult to place. Despite the supposition that the middle and upper Pennsylvanian formations are missing there is little evidence of an unconformity or disconformity between the two formations.

A petrographic study was made of three sections along the northern boundary of the Quadrant and Phosphoria basin. Samples were collected from the Quadrant and up through the lowermost phosphate beds in the Phosphoria. An effort was made to establish the environmental background leading up to the deposition of the phosphate beds and to determine the lithologic differences across the paleontological boundary.

* This work was done in 1949–1950 at the Pennsylvania State University.

During the summer of 1948, 130 samples were collected in southwest Montana across the Quadrant-Phosphoria boundary from three trenches dug by members of the U.S. Geological Survey.

The location of the three trenches is shown on the map (fig. 1). They lie roughly in a straight line, 50 miles long, trending northwest. The stratigraphic sections showing where samples were collected are shown in figure 2. At Wadham's Spring 70 feet of Quadrant and 55 feet of Phosphoria were sampled; at Kelly Gulch 25 feet of Quadrant and 115 feet of Phosphoria were sampled; at Sheep Creek five feet of Quadrant and 105 feet of Phosphoria were sampled (fig. 2).

The trenches were all dug on the limbs of moderately folded anticlines which dipped 40° to 50°. The folding at Kelly Gulch was more intense than at the other two localities. The entire region contains numerous faults and small igneous dikes.

Stratigraphy

The Quadrant formation, although having few fossils, is thought from its known faunal and stratigraphic relations to be of lower Pennsylvanian, Des Moines age (Branson, 1939).

Scott (1935) stated that "the Quadrant quartzite is unquestionably a westward extension of the Tensleep sandstone." The uniform Tensleep sandstone is found in practically all of Wyoming, but it wedges out to the north and northwest in Montana where it is known as the Quadrant. It grades southward into the Weber of Colorado and Utah and westward it is correlated with the lower portion of the Wells (Branson, 1939).

The Tensleep in central Wyoming, near the type area, is 200 to 500 feet thick and composed largely of sandstone cross-bedded in the upper part. Massive dolomite beds are abundant in the upper part and limestone in the lower; shale is seldom found; chert is abundant as nodules and nodular beds. Many of the sandstone beds are cemented by carbonates. The Quadrant in southwest Montana is a uniform, fine-grained, slightly cross-bedded sandstone similar to the Tensleep. The formation ranges from 75 to 1000 feet thick, and is usually between 200 and 300 feet thick.

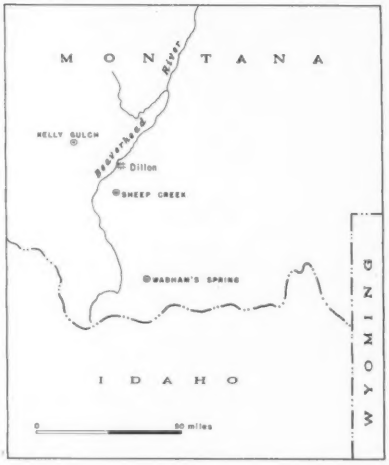

FIG. 1.—Map showing the location of the three sections sampled.

Rocks of the Phosphoria formation are found at scattered areas in central and northern Nevada and Utah, central and western Wyoming, and most of eastern Idaho and western Montana.

Frenzel and Mundorff (1942) reported the first discovery of fusulinids in the Phosphoria, found near Three Forks, Montana, at the base of the Phosphoria, silicified within chert nodules. They have been identified as *Schwagerina laxissima* and a new species named *Pseudoschwagerina montanensis*. *Schwagerina laxissima* is a zonal guide fossil found in the basal Permian (Wolfcamp and Heuco) of Texas. Miller and Cline (1934) reported an occurrence of ammonites in the Sublette Mountains of western Wyoming, 60 feet above the base of the Rex chert. The ammonites, according to Miller and Furnish (1940), indicate Word or possibly Leonard age for the enclosing rocks.

Thus, faunal evidence suggests that beds of the Phosphoria formation represent more than one epoch of the Permian.

The Phosphoria formation, 460 feet thick at its type section (Richards and Mansfield, 1912) in southeastern Idaho, is divided into two distinct lithologic units. The upper unit consists of massive limestones and cherts and is called the Rex chert member. The lower unit comprises the Phosphatic shales.

The Rex chert member is conspicuous for its massive beds of chert which reach a thickness of 60 feet. The Rex member also contains beds of cherty limestone and shale—in fact, these are the dominant components in places.

The Phosphoria formation, though there is a great deal of minor variation over short distances, has a fairly uniform lithology and thickness through southeast Idaho, northern Utah, and southwest Wyoming. The thickness of the

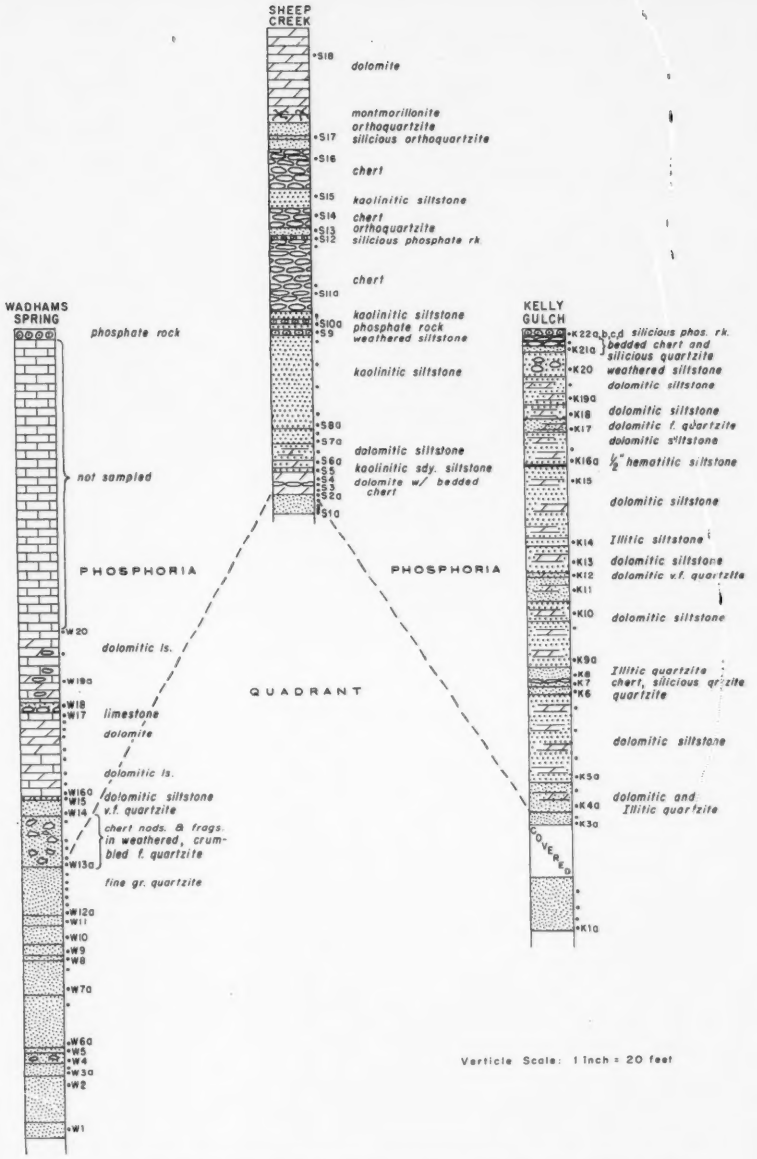

FIG. 2.—Stratigraphic section at the three trenches. Dots show location of samples.

Phosphatic shale member is generally around 180 feet (McKelvey, 1946).

In central and northwest Wyoming, where the formation is 100 to 300 feet thick, the phosphate rock occurs at two principal horizons. One is near the base of the Phosphoria formation, at times lying directly on top of the Tensleep, and the second is 20 to 130 feet above the lower zone. The maximum thickness of the upper bed is 50 feet and it thins to a few inches in the northwest corner of Yellowstone Park.

The two zones are separated by chert, limestones, and shales. The lower zone is thinner and not as extensive as the upper zone. It is also different from the other in that it consists at some places almost entirely of detrital phosphatized mollusk shells. This lower bed is usually separated from the underlying Tensleep or Quadrant by limestones and cherts. The underlying limestones are sometimes grouped with the Quadrant. The upper portion of the Phosphoria, equivalent to the Rex chert of Idaho, is composed largely of quartzite in western Wyoming and grades into limestone in central Wyoming.

In west central Montana and southwestern Montana the Phosphoria varies from 100 to 500 feet in thickness; it is similar to the Phosphoria of western Wyoming. The upper oolitic phosphate zone is everywhere present, varying in thickness from eight to 20 feet. The oolitic phosphate rock occurs as individual layers an inch to a foot or two in thickness and has five to seven or more members; they are separated by dark phosphatic shales.

Cherts, limestones, dolomites, and dolomitic, kaolinitic, and illitic siltstones separate the two zones. The lower zone is absent in north central Montana, but it is present in southwest Montana. The zone is generally quite thin, and consists of two or three one inch bands of phosphate rock separated by a few inches of phosphatic chert. West of Lima this zone is 50 feet thick (Cressman, 1950). This lower zone lies from 30 to 150 feet above the Quadrant quartzite, separated from it by dolomites, cherts, siltstones, and, occasionally, thin beds of quartzite. Richards and Pardee (1925) found places in the Melrose district where the lower phosphate bed rested unconformably upon the Quadrant quartzite; however, in the region sampled by the writer there was no evidence of an unconformity.

The upper or Rex chert member, in southwest Montana, is a massive, bedded chert.

MINERALOGY AND COMPOSITION

Major Constituents

Quartz

Detrital quartz is abundant in both the Phosphoria and Quadrant formations of southwestern Montana. The orthoquartzites, which are the predominant rock type of the Quadrant formation, contain about 80 percent to 97 percent detrital quartz. Orthoquartzites form only about eight percent of the Phosphoria formation but detrital quartz occurs in every rock type in the Phosphoria. It makes up about 45 to 85 percent of the siltstones, two to 40 percent of the dolomites and limestones; at least a few grains are generally present in the cherts.

Plutonic Quartz.—In this group are placed grains which have no or moderate strain shadows and relatively few inclusions.

By far the largest portion (about 95 percent) of the detrital quartz in both formations was derived originally from plutonic quartz. Though some of the grains are free of inclusions, most contain bubbles and vacuoles (cavities) of various types: (1) small disseminated bubbles, (2) bubble trains or planes, (3) elongated and dendritic vacuoles. All three types occur in both the Quadrant and Phosphoria.

The disseminated bubbles are the most abundant, frequently being the only

inclusions in a grain. Bubble planes that cross one another are occasionally found in the two formations but they never cross grain boundaries, indicating that they were formed before the grains were deposited as Quadrant and Phosphoria quartzites.

Bubble planes seldom occur alone; they are usually associated with disseminated bubbles, and occasionally with the elongated and dendritic type of cavities.

The elongated and dendritic type of cavities, fairly abundant in most of the Quadrant samples, are rare in the Phosphoria samples.

On a quantitative basis, bubble inclusions are more abundant in the Quadrant than in the Phosphoria samples, mainly because bubbles appear to be much less prevalent in silt-sized (most abundant in the Phosphoria) than in sand-sized particles. This is probably because of the tendency of quartz to fracture along the direction of planes of bubbles. The number of bubbles in samples of equivalent grain size is about the same in both formations.

Crystal inclusions in both formations are identical in type and similar in abundance. Rutile needles are the most common, occurring in as many as 10 to 15 grains in a normal thin section. Brown, olive, and gray tourmaline, colorless zircon, apatite, and quartz are present as round, prismatic inclusions.

At Wadham's Spring and Sheep Creek about 30 percent of the plutonic quartz grains in both formations show straight extinction, and the other 70 percent has from slight to moderate undulose extinction. Some of the undulose extinction may be due to post-diagenetic folding, as at Kelly Gulch where all of the quartz grains show moderate to strong undulose extinction and have moderately crenulated borders. (Kelly Gulch is in the Pioneer Mountain Range where the sediments are more intensely folded than at the other two localities.)

End Phase Quartz.—All quartz which crystallized in the late stages of magmatic differentiation, including pegmatitic and vein quartz, is considered end phase quartz. This quartz is characterized by flamboyant extinction, comb structure, and an abundance of parallel bubble planes.

End phase quartz is present in all samples of the Quadrant in amounts ranging from two to four percent. Traces occur in the Phosphoria rocks.

Metamorphic Quartz.—The metamorphic quartz grains are composed of tightly packed individual grains which range in size from 10 to 50 microns, and whose borders show various degrees of crenulation. The extinction in all cases is undulose and in some grains anomalous blue interference colors can be seen. A few bubbles may be present, and occasionally inclusions of sillimanite may be seen.

This type of quartz is found in all samples of the Quadrant in amounts ranging from one to two percent. Traces are found in the Phosphoria quartzites.

FIG. 3.—(S6b) Lath-shaped quartz as a constituent of dolomitic siltstone. ×50.

Volcanic Quartz.—Several Phosphoria silt and sandstone samples contain four to 10 percent of lath shaped quartz grains (fig. 3). Most of the laths consist of single grains, but many are made up of several individual quartz units. Ray (1947) and Wager, et al (1953), have reported lath quartz of this nature paramorphic after tridymite.

Authigenic Quartz Overgrowths.—Authigenic quartz overgrowths are present in all samples of the Quadrant and in all of the orthoquartzite samples and one dolomitic limestone sample of the Phosphoria. The percentage present varies from less than one to 10 percent and averages one to two percent of the total quartz. With the exception of two samples, overgrowths occur on 65 to 90 percent of the grains in the Quadrant quartzites and generally occur on only a few grains in the Phosphoria samples, though in one sample they occur in 30 percent of the grains.

Though there was an abundance of silica in the Phosphoria sea, as indicated by the large amount of chert deposited, few overgrowths formed. In nearly all cases the quartz sand and silt rocks contain authigenic kaolinite and illite in the interstices. The chemical environment was evidently more favorable for the growth of clay than for the development of quartz overgrowths.

The overgrowths, in optical continuity with the quartz, are colorless, and frequently contain bubble inclusions. They average 10 to 15 microns in thickness and reach a maximum of 50 microns.

The overgrowths, in all cases, in both formations are separated from the detrital grains by a single layer of bubbles. These bubbles have all the characteristics of the bubble inclusions within the detrital quartz grain; they are one layer in thickness, parallel to the quartz grains and are narrow and elongated, varying from one to 10 microns in size. A possible explanation is that the original quartz grains were pitted or corroded and the silica in solution was first deposited on the elevated ridges on the grain surface and gradually coalesced, forming a solid sheet, as the hygroscopic silica held the water in the depressed areas.

Irregular patches of microcrystalline quartz (chert) occur locally in both formations as matrix. When this is present the oriented quartz overgrowths grade into the unoriented grains of microcrystalline quartz. The microcrystalline quartz shows evidence of originally replacing a fossiliferous calcite cement.

Microcrystalline Quartz.—Microcrystalline quartz comprises approximately 95 percent of the chert rocks and grains in the Quadrant and Phosphoria formations. It occurs in two ways: (1) as detrital, second cycle grains, and (2) as nodules and beds and as a cement in quartzites. The first type will be described in detail under rock fragments.

Round, elliptical chert nodules, ranging from five to 20 mm, are commonly present in the carbonate rocks. The nodules are composed largely of microcrystalline quartz, though fibrous chal-

FIG. 4.—(W13b) Electron micrographs of chert showing irregular xenomorphic patches of microcrystalline quartz. ×1530.

cedony frequently comprises 10 percent of the chert. The nodules have a sharp contact with the enclosing carbonate rock.

Under the petrographic microscope microcrystalline quartz occurs as tightly packed grains of quartz, varying in size from less than one to 60 microns in diameter, and averaging five to 10 microns. The smaller grains appear to have irregular interpenetrating boundaries, (electron-micrographs show that in many cases this is merely a pseudo feature as the chert can be seen to be composed of small grains which have sharp, distinct borders) whereas the larger grains have, as a whole, smooth, sharp boundaries. The large majority of the grains are xenomorphic and subequant; a few of the larger grains are hypidiomorphic. All of the grains are anisotropic, though some of the smaller grains must be examined under extremely brilliant light in order to see any interference colors. All of the grains show a pseudo-undulose extinction.

Inclusions are common, with bubble inclusion being the most abundant. Fine shreds of sericite or illite are common, though it is not clear whether they occur within the grains or between the grains. Irregular chunks of clay are sometimes observed. In many grains carbonate inclusions are abundant, these frequently show indistinct outlines of replaced fossils. Dolocasts and dolomite rhombs are common.

The color of the quartz crystals varies from colorless (rare) to various shades of tan and dark gray. The tan color is due to bubble inclusions and the gray color is probably due to inclusions of clay or organic matter. The index of refraction in the clear colorless varieties of microcrystalline quartz is equal to that of igneous quartz, whereas microcrystalline quartz containing a considerable number of bubble inclusions has an index lower than quartz. This is due to the effect on the refraction of light caused by the bubbles, which have an index considerably less than quartz.

FIG. 5.—(K20) Electron micrograph of chert showing distinct grain outlines. Grains marked with an A appear to have been fractured. ×2400.

Electron micrographs were taken of collodion replicas of four chert nodules from the Phosphoria. As can be seen by the prints (figs. 4 and 5), there is a striking difference in the electron micrographs of two of the samples—both chert nodules. Sample W13b (fig. 4) is composed of indistinct, tightly packed, xenomorphic patches of quartz; sample K20 (fig. 5) is composed of distinct, tightly packed, individual particles.

Figure 5 is a low power electron micrograph showing a relatively large, typical field of view of chert sample K20; the grains vary from equant to prismatic, the majority average one to three microns in size, and reach a maximum of 10 microns. Most of the grains have smooth continuous surfaces that appear to be normal grain surfaces. Several larger grains appear to have been broken across.

The pseudo-undulose extinction of microcrystalline quartz can be easily explained now that the grain size can be determined. If the average grain size is three microns, a thin section 30 microns

thick will have 10 layers of microcrystalline quartz particles atop one another; any variation in orientation would give the effect, in the thin section, of undulose extinction.

The chert appears to be of two distinct types under the petrographic microscope. In the thin section of sample W13b the chert is extremely fine-grained, averaging one micron, and has a typical pepper and salt birefringence.(The grain size of chert in thin sections is based on a measurement of distinct birefringent units and is not an actual measure of grain size.) The thin section contains one to three percent of finely disseminated carbonate grains, less than one micron in diameter. There are also fusulinids and other fossil fragments which have their calcite shells replaced by microcrystalline quartz. It is possible that, in addition to the fossils being replaced, a limestone matrix containing the fossils has been replaced by microcrystalline quartz, and the disseminated carbonate grains are a remnant of the original limestone.

The presence of dolomite in the outer borders of chert nodules and the presence of replaced calcite fossils in the interior of the nodule (assuming that dolomitization destroys calcite fossils) means that the time of replacement by microcrystalline quartz and by dolomite overlap. The microcrystalline quartz replaced some of the calcite fossils before they were destroyed by dolomitization, but as replacement by the microcrystalline quartz continued dolomite replaced the surrounding calcite, and the microcrystalline quartz, in turn, replaced the dolomite.

Sample K20 is coarser grained than sample W13b (under the petrographic microscope), the grains averaging five microns. There is only a trace of disseminated carbonate grains less than one micron size and no fossils. There are many lenses and veins of coarse grained vein quartz and fibrous chalcedony in the chert. It appears that the chert was originally deposited as a primary gel and that upon consolidation dehydration

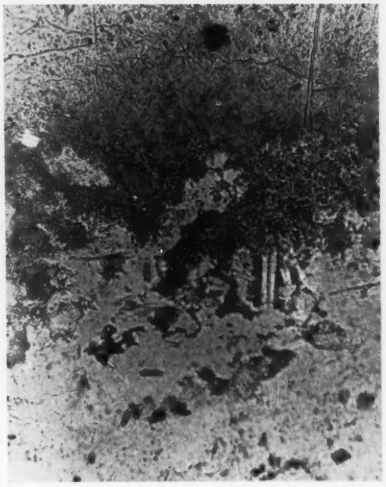

FIG. 6.—(K22a) Interfingering microcrystalline quartz and collophane. The collophane apparently replaced the quartz. The darker colored material is the collophane. ×70.

cracks formed; a reorganization of the silica in the gel, or the addition of new silica from some extraneous source, formed the larger vein crystals and the fibrous chalcedony in the dehydration cracks. If this chert sample (K20) were formed by primary deposition and the other sample (W13b) formed by replacement it would give a ready explanation of the differences apparent in the electron micrographs.

Thin beds of chert are commonly found intimately associated with phosphate rock and as a result they commonly contain a few grains of glauconite and collophane. Sample K22a shows the massive, authigenic collophane interfingering with microcrystalline quartz (fig. 6), the collophane appears to have replaced the chert since spots can be seen where it corrodes and partially replaces grains of detrital quartz and chert.

At Sheep Creek a 17-foot bed of chert separates two thin beds of phosphate rock and a nine-foot bed of chert occurs

above the upper phosphate bed. Siliceous sponge spicules are abundant; carbonate grains are rare; five to 10 percent of silt size detrital quartz and one to two percent of muscovite and illite are commonly found in the chert. Most of the bedded chert has an interfingering white and gray color. The white chert is composed of coarse-grained, microcrystalline quartz and chalcedony, whereas the gray chert is composed of fine-grained (one micron) microcrystalline quartz, which generally contains more detrital material than the white chert.

Chalcedony.—Chalcedony is a finely fibrous variety of quartz that is generally associated with microcrystalline quartz in chert. It occurs both in detrital grains of chert and in authigenic chert. Detrital grains are scarce, but a few were found in all of the orthoquartzite and sandy carbonate samples. In addition to forming the entire grain, it frequently is found as veins crossing grains of microcrystalline quartz. The mineral generally has a fibrous extinction and is length-fast; however, in sample S12, spherical grains have a feathery, flame-like birefringence. These grains (S12) have a concentric structure which can be seen in reflected light; they commonly contain altered pyrite cubes.

Chalcedony forms approximately five percent by volume of the authigenic chert examined. It is generally less than this but on rare occasions forms as high as 30 percent of the chert where fossils are abundant. Occasionally it is found as spherulites with the typical black cross showing under crossed nicols; more commonly it is associated with replaced fossils, occurring as short clear fibers replacing the fossil shell and as long bubble-filled fibers (which average 25 microns in length and range up to 150 microns) filling the inside of the shell. It is also found as irregular patches growing from a nucleus. The nucleus is microcrystalline quartz which grades into coarse vein quartz and from that into fibrous chalcedony.

Rock Fragments

Only quartzite and chert are present as rock fragments.

Metamorphic Quartzites.—Described under quartz (metamorphic).

Chert.—Detrital chert grains are composed largely of the previously described microcrystalline quartz. In a few grains small veins of chalcedony cross grains of microcrystalline quartz. There are one or two grains composed completely of fibrous chalcedony in each sample.

The chert grains have a range in size similar to that of the quartz grains. The chert is usually better rounded than the quartz, has smooth borders, and is slightly elongated. Detrital chert is present in all of the orthoquartzites in amounts ranging from one to 11 percent and averages four percent in both formations.

The chert can be subdivided on the basis of the size of the microcrystalline quartz making up the grain. There are three natural size groups: (1) grains less than two microns—mere pinpoints that are dark under crossed nicols, (2) grains between five and 15 microns, (3) grains between 15 and 60 microns. Mixtures of two or more types are not common except when the grains are crossed by veins. The three types occur in the same relative amounts in nearly all samples of both formations. The largest size occurs only as a trace—frequently some of the crystals in this size are singly terminated and hypidiomorphic. The second type (5–15 microns) is generally the most abundant, making up 65 to 75 percent of the total chert in a sample. The first type (less than two microns) is next in abundance, making up 25 to 35 percent of the total chert. In only one sample, W5, is the relative abundance of the two major types reversed.

Small carbonate inclusions and dolomite rhombs are infrequent. Sericite-illite shreds, five to 10 microns long, are common as inclusions in the chert grains; they are unoriented and vary from one or

two per grain, up to three percent of the total grain; they are found in the chert of all samples, generally occurring in approximately 25 percent of the chert grains; they commonly occur in the finer grained chert.

Micas

Muscovite.—Muscovite shreds are scarce in the orthoquartzite, and are most abundant (0.5 to two percent) in the siltstones and impure carbonate rocks. In a thin section they show up as typical, long narrow shreds, frequently slightly bent as they are wrapped around quartz grains. The shreds range from 0.03 to one mm long and are 0.005 to 0.025 mm wide. The distinctive round oval shape of these flakes, which can be seen in the heavy mineral separate (described under heavy minerals), is not evident in any of the thin sections.

Chlorite.—Chlorite is found only in the slightly metamorphosed orthoquartzite at Kelly Gulch. It occurs as small shreds, average 0.02 mm intimately mixed with limonite as a coating on quartz grains, and as aggregates of books filling voids between grains. It forms one to two percent of the rock.

As kaolinite is found in the unmetamorphosed orthoquartzites from the other two localities it seems likely that the chlorite was formed by low-grade anamorphism of kaolinite.

Biotite.—Several flakes of altered biotite were found in sample K17. It occurs as subequant, angular flakes. About two-thirds of the flakes consist of colorless, roughly parallel shreds. The brown portion of the flake has a light brown to dark brown pleochroism. The flakes apparently have had much of the iron removed from them.

Collophane

Collophane occurs as individual grains and as cement. It is concentrated in thin bands, a few inches wide, separated by beds of orthoquartzite, limestone, and chert, and forms the lower phosphate zone in the Phosphoria formation. At Sheep Creek, twenty feet above this zone, there is a thin (1.4 mm) band of collophane cement. In addition, a few grains of collophane are found in carbonate rocks both above and below the phosphate zone; collophane bone fragments were found in one sample of the Quadrant orthoquartzite.

The collophane varies from colorless to medium brown. It is generally isotropic, but frequently bone fragments have narrow bands which show a gray to white birefringence. The collophane consists of a submicroscopic aggregate of apatite crystals which contain varying amounts of adsorbed and capillary water (Frondel, 1943). X-ray patterns show that the collophane in the sample from the lower phosphate zone is composed of fluorapatite.

Electron micrographs of a crushed sample of oolitic collophane show highly irregular, indented borders which are presumably caused by the aggregate (apatite) structure of collophane.

The collophane grains can be divided into three types: (1) bone fragments, (2) oolites, and (3) irregular grains. The bone fragments have a great variety of shapes with their borders ranging from angular to round. They frequently show some weak birefringence. The oolites are spherical to egg-shaped with round borders. They range in size from 0.14 to 0.90 mm. The oolites have one or two concentric bands. Inclusions are abundant: quartz, chert, collophane, and glauconite. Oolites are found only in the phosphate zone. The third or irregular grain type is the only type found in the carbonate rocks. The grains are generally irregular with rough borders though occasionally grains are placed in this type with round borders that do not appear to fall into either of the other two types. The irregular patches sometimes appear as wormy intergrowths in calcite grains, which indicates replacement of the calcite by the collophane. Inclusions of quartz grains and mica shreds are present in this type.

Collophane rock contains collophane which acts as a cement and shows no grain boundaries; it is found commonly in the phosphate zone. The cement forms 50 to 60 percent of the rock, enclosing grains of quartz, chert, mucovite, glauconite, and collophane (generally oolitic). The cement is generally brown. At Sheep Creek it is structureless; at Kelly Gulch it has irregular streaks which show a gray to white birefringence (this may be due to strain which was greater at Kelly Gulch than at Sheep Creek). The colophane cement frequently intimately interfingers with microcrystalline quartz (fig. 6). Outlines of sponge spicules are found in both the collophane and the microcrystalline quartz. Detrital chert, enclosed in the collophane cement contains irregular patches of collophane and occasionally has irregular replacement borders. From the above evidence it appears possible that some of the collophane has replaced microcrystalline quartz.

Clay Minerals

Three clay minerals were identified: kaolinite, illite, and montmorillonite. The identification was verified, in most cases, by X-ray patterns, though differential thermal patterns and electron-micrographs were also used. Some clay is found in all rocks. Table 1 shows the type of clay found in each sample and the relative abundance based on the intensity of the X-ray pattern.

A trace of kaolinite is found in all the Quadrant quartzites except at Kelly Gulch where it is believed to have altered to chlorite. Kaolinite is present in the Phosphoria at Wadham's Spring and at Sheep Creek and occurs in a few samples at Kelly Gulch; illite is the most abundant clay at Kelly Gulch but is found only in trace amounts at the other two trenches; montmorillonite is found in one sample at Wadham's Spring and one at Sheep Creek.

Kaolinite.—Kaolinite is widely distributed in the Quadrant, though comprising less than one percent of the rock. It commonly occurs as linings of voids, sometimes completely filling the voids; less commonly it is found between grains.

The detrital silicate rocks in the Phosphoria of Wadham's Spring and Sheep Creek contain amounts of kaolinite varying from one to 20 percent, based on estimates from thin sections. About half of the thin sections contain one to five percent and the other half 15 to 20 percent kaolinite. The finer-grained rocks (siltstones) generally contain the larger amounts of kaolinite. The clay is unoriented, and where it is present in small amounts it commonly occurs as fine shreds two to five microns long coating grains. Equant kaolinite books, worms, and irregular masses, averaging 0.08 mm in width, are present in all samples; the books generally contain some illite interlayers which have a higher birefringence than the kaolinite. In samples where there is 15 to 20 percent of clay, kaolinite occurs as a nest between grains, frequently forming a roughly lenslike mass. Some of the kaolinite is probably authigenic.

Figure 7 is an electron micrograph of the insoluble residue of dolomite sample S4a which gave a strong kaolinite X-ray pattern. This micrograph shows typical well-developed, hexagonal kaolinite flakes which range in size from one-half to four microns.

Illite.—A few shreds of illite are present in the kaolinite clay found in the Quadrant and in the Phosphoria of Wadham's Spring and Sheep Creek. At Kelly Gulch illite is the most abundant clay mineral.

Illite is present in amounts ranging from one to 20 percent based on estimates from thin sections. It commonly forms thin coatings between quartz grains. Frequently it is found replacing dolomite and quartz grains. Nests of illite books and vermicular grains, averaging 0.04 and 0.08 mm, are common (fig. 8). The illite appears to be largely authigenic.

TABLE 1.—*X-ray Data*

	Qtz.	Dol.	Cal.	Kao.	Ill.	% Sol. in HCl	
Watham Springs							
W1	M			M			2μ
W3a	M			M			
W9	M			W			
W12a	M			M			
W12d	M			M			
Quadrant Phosphoria							
W15	M	M		M		37.0	BULK
W16b		10	85			94.5	
16c		48	48			96.4	
16d		49	49			97.8	
16e		97				97.3	
16f		98				98.0	
16g	S	63		M		63.0	
16h		92				91.6	
W17	W		48			88.0	
W18	S						
W19a	W	43	44			87.2	
19b		45	45			90.0	
W20	W	41	41			81.6	
W15	M			M			RESIDUE
W16b	M			M			
16c	M			M			
16d	S			M			
16e	S			W			
16g	M			M	W		
16h	M						
W17	S						
W19a	M						
19b	M			Montmorillonite			
W20	S			W			
Sheep Creek							
S1a	M			S			31.5μ
S2a	M						BULK
2b		85				85.3	
S3		92				91.8	
S4		9				92.6	
4a				M			
S5	S			W			
S6a	W	69				68.8	
6b	S	33		M		32.8	
S7a	S			M			
7b	S			M			
S8a	S			M			
8b	S			M			
8c	S			M			
8cc	S			M			
8d	S			S	W		
S9	Fluor Apatites			(S)			
S10a	S			S			
10c	S			S			
S11c	S						

TABLE 1.—*X-ray Data (Con't)*

	Qtz.	Dol.	Cal.	Kao.	Ill.	% Sol. in HCl	
			Sheep Creek				
S12	M F. Apatite			(S)W			
S15	S						
S17	S						
S18		S				97.4	
S2b	M			M			RESIDUE
S3	M			S			
S6a	S			S			
S18	S						
S18	W Montmorillonite				(S)		
			Kelly Gulch				
K4a	S	20			W	20.2	BULK
4b	S	24				24.0	
K5a	S	15			M	14.5	
5b	M	15			W	15.3	
5c	S	15			W		
5d	S	15		W			
K8	S				W		
K9a	S	19				18.5	
9b	S	21		W	W	21.2	
K10	S	21		W	W	21.2	
K11	S	17			W	16.5	
K12	S	16			W	16.0	
K13	S	16			M	15.8	
K14	S				S		
K15	S	23			M	22.7	
K16a	M	15			M	14.9	
16b	S	16				16.0	
K17	S	16		W	W	15.8	
K18	S	21			W	20.6	
K19a	S	16			M	15.8	
K19b	S	21			W	21.0	
K22a	S	Fluor Apatite			(S)		

S = Strong.
M = Moderate.
W = Weak

The table lists the intensity of the X-ray spectrometer patterns of minerals in the bulk sample, size fractions, and residues of most of the samples collected from two formation.

Figure 9 is an electron micrograph of the fine fractions of an illitic siltstone from Kelly Gulch. It shows xenomorphic, thin, smooth bordered flakes, typical of the better developed illites.

Montmorillonite.—Two occurrences of montmorillonite were found. Sample S18 was taken from a network of thin veins, one to two inches wide, in a thick bed of dolomite at Sheep Creek. The veins, which bear no relation to bedding, consist of dark brown, hard, almost pure montmorillonite clay. X-ray patterns indicate that approximately 10 percent quartz is associated with the montmorillonite.

The second occurrence is in the insoluble residue (10 percent) of the dolomitic limestone at W19b. The montmorillonite occurs, entrapped in, and coating, authigenic microcrystalline quartz.

Though some of the clay is detrital, the many books, nests, and vermicular

worms of kaolinite and illite indicate that much of it is probably authigenic. The fact that different clays are concentrated in different areas indicates that growth was controlled by the local physical and chemical environments.

The vein montmorillonite in sample S18 is likely of hydrothermal origin.

Hematite and Limonite

Hematite and limonite occur together, the limonite generally weathering from the hematite. These iron oxides are found in nearly all samples of the Phosphoria and approximately in half of the samples of the Quadrant.

Limonite is found coating grains, lining voids, intermixed with various clays, and as irregular patches, veins, and stringers. Frequently it can be observed spreading outward in irregular stringers from isolated crystals, masses, or veins of hematite.

Hematite, as mentioned, is generally altered, completely or in part, to limonite. At Kelly Gulch sample K14 contains a thick band, one-half inch wide, and several smaller stringers where hematite occurs as translucent, golden red, crystalline flakes. Quantitative chemical tests show that this band contains 47 percent iron oxides. Hematite in most of the samples is found as xenomorphic, isolated grains or narrow veins.

In chert, limonite and hematite are frequently found as pseudomorphs after pyrite.

FIG. 7.—(S4a) Electron micrograph of kaolinite in the insoluble residue of dolomite. ×7000.

Carbonates

Two carbonate minerals, calcite and dolomite, are abundant in the Phosphoria formation, but are found in only four samples of the Quadrant (and then only as traces in overgrowths or in microcrystalline quartz cement).

Table 1 shows the percentage of total carbonate minerals present in each sample, based on the percentage by weight of the amount of each sample soluble in 1 N hydrochloric acid. This table also indicates the percentages of calcite and dolomite present. These percentages were based on the relative intensity of the X-ray reflections from the major calcite and dolomite peaks. Measured intensities of standard calcite-dolomite mixtures indicate that the percentages are correct to approximately ± five percent.

Calcite.—Most of the calcite occurs in the Wadham's Spring samples. In only one sample, W17, is calcite found without dolomite being present.

The calcite grains are colorless, subequant, and generally have jagged, irregularly interfingering borders; infrequently xenomorphic grains occur with smooth, straight borders.

The grains average 0.35 to 0.40 mm in diameter and range from 0.14 to 0.8 mm in diameter. Generally the larger calcite grains contain inclusions of unoriented calcite particles. These inclusions are gray in color, roughly equant, and average five to 100 microns. The calcite inclusions comprise 30 to 70 percent of the large calcite grains, are evenly distributed, and are seldom in contact with one another.

Extremely fine-grained (one to three microns) calcite is found in chert showing partially replaced fossil outlines.

FIG. 8.—(K8) Nest of antigenic illite in a Phosphoria quartzite sample. ×220.

Dolomite.—In approximately half of the carbonate rocks at Wadham's Spring, dolomite occurs as the only carbonate mineral; in the other half it occurs associated with calcite. At Sheep Creek and Kelly Gulch dolomite is the only carbonate mineral present. Where dolomite is associated with calcite it forms distinct hypidiomorphic to idiomorphic inclusions in the larger calcite grains.

In the dolomite rock the dolomite grains are colorless, equant, have distinct, smooth grain boundaries, and are generally xenomorphic to hypidiomorphic in shape. The grains with an idiomorphic tendency commonly show rhombic cleavage. The grains range in size from three to 50 microns, with the great majority, in the carbonate rock, being in the 15 to 30 microns range; where the grains occur as a cement most of the grains fall in a three to eight micron range.

Where dolomite occurs as a cement (Kelly Gulch) it is found as isolated grains and stringers, and very seldom are grains welded together in the fashion of a conventional chemical cement.

The irregular gradation of dolomite into calcite, the occurrence of isolated dolomite grains in calcite, and the absence of fossils in the dolomite indicate that the dolomite has probably replaced the original calcite. The presence of dolomite breccia in the dolomitic limestones suggests that dolomitization had taken place relatively soon after precipitation of the primary calcite when the newly formed dolomite was within the sphere of wave action.

Insoluble Residues.—The insoluble residues of all carbonate rocks (greater than 50 percent carbonate minerals) were examined. Several samples contained as much as 40 percent insoluble residue, but the majority contained from two to 18 percent residue, averaging 11.2 percent. The residue consists of highly etched, detrital quartz grains (average grain size around 0.03 mm), authigenic chert in the form of irregular aggregates of microcrystalline quartz (0.04 to 0.3 mm), kaolinite, illite, muscovite, microcline,

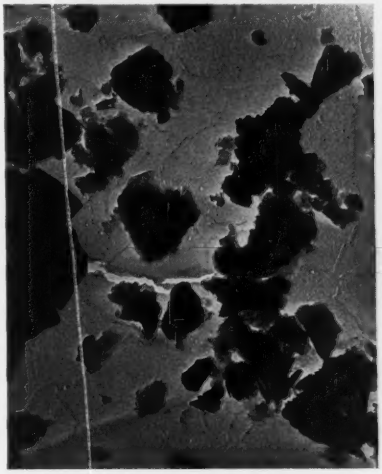

FIG. 9.—(K14) Electron micrograph of illite from a dolomitic siltstone. ×6800.

zircon, tourmaline, and rutile. At Wadham's Spring the chert averages 26 percent of the residue; at Sheep Creek, 74 percent.

PETROGRAPHY

Only eight samples were suitable for mechanical analyses: six from the Quadrant formation and two from the Phosphoria formation. U.S. Standard sieves were used.

The Quadrant is typically a fine-grained sandstone. The six samples contain an average of 54.5 percent fine sand, which is the predominant size in five samples. In W3a, however, very fine sand predominates. The very fine sand is second in abundance, averaging 25.8 percent of the total detrital grains; medium sand is relatively scarce, forming 6.5 percent of the total; coarse sand is entirely lacking. The silt (62.5–3.9 microns) and clay (less than 3.9 microns) fractions together average 13.2 percent of the total weight as computed from the mechanical analysis. Pipette analyses show that the clay fraction constitutes approximately two percent of the bulk sample, thus leaving 11 percent of the material in the silt size. The clay fraction is shown by X-ray analysis to consist largely of quartz, with minor amounts of kaolinite; some of the quartz was possibly produced during the mechanical disaggregation of the sand.

In the two Phosphoria samples there is no striking modal class. The medium sand constitutes 4.5 percent, the fine sand forms an average of 35.5 percent, the very fine sand forms an average of 31.1 percent, and the silt and clay class averages 27.0 percent of the total. The real differences between the two formations is due to a slight increase in the amount of the very fine sand class, and a much larger increase in the silt and clay class of the Phosphoria samples.

The cumulative curves (fig. 10) are similar to those of other well-sorted orthoquartzites and to those of modern near-shore sands. The amount of material finer than 4.5ϕ in the Quadrant is less than 10 percent in all samples except W9 where it is 16 percent. The Phosphoria samples have 21 and 22 percent material finer than 4.5ϕ.

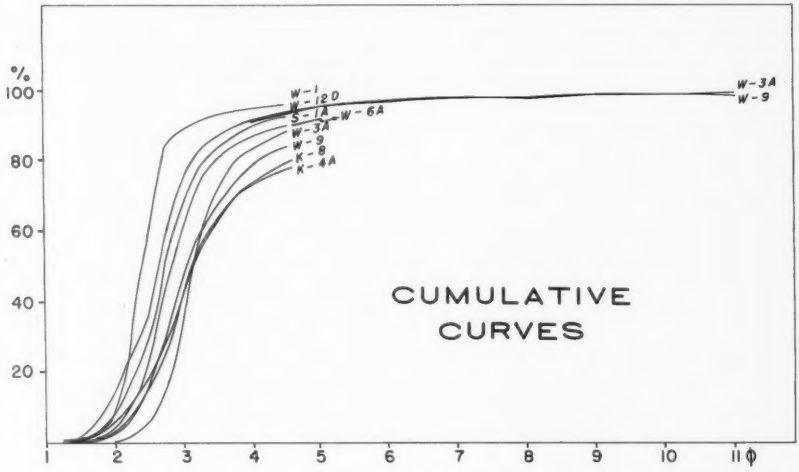

FIG. 10.—Cumulative curves of six Quadrant and two Phosphoria (K4a and K8) samples.

TABLE 2.—*Significant size parameters*

Sample	Md	Mo	So	Sk	QDϕ	PDϕ	Mdϕ	Moϕ
				Phosphoria				
K8	.114	.129	1.60	.66	.68	1.58*	3.13	2.95
K4a	.116	.129	1.70	.66	.75	1.65*	3.10	2.95
				Quadrant				
W12d	.159	.159	1.33	1.00	.40	1.00	2.65	2.65
W9	.129	.149	1.50	.69	.58	1.25*	2.95	2.75
W6a	.139	.149	1.31	.92	.39	1.13	2.85	2.75
W3a	.112	.116	1.29	.89	.38	1.00	3.15	3.10
W1	.189	.196	1.26	.85	.33	.50	2.40	2.35
S1a	.151	.159	1.34	.93	.39	.98	2.73	2.65

* Extrapolated.

The P_{90} value could be read from the cumulative curves of four of the samples, but it was necessary to plot the grain size analyses of samples W9, K4a and K8 on arithmetic probability paper and to extrapolate in order to obtain their P_{90} values.

The sorting value, So (in mm) was computed using the first and third quartile taken from the cumulative curves. Table 2 contains a list of the So values which range from 1.26 to 1.50 for the Quadrant and are 1.60 and 1.70 in the Phosphoria. All values are less than the 2.5 values given by Trask (1932) as the maximum value for well-sorted sediments. They also fall within the one to two range given by Hough (1940) for the sorting value of near-shore marine sediments. The Phosphoria sandstones fall in the well-sorted range, though they are slightly less well-sorted than the Quadrant.

The QDϕ and PDϕ values $(Q_3-Q_1/2)$ and $(P_{90}-P_{10}/2)$ which measure the quartile and percentile deviation in ϕ units are a direct measure of sorting. The QDϕ is the ϕ value equivalent to the So value. As would be expected, these values (table 2) show the same relationship between formations as the So values.

The average of the Md value of the six Quadrant samples is 0.146 mm; the average of the Md value of the two Phosphoria samples is .115 mm.

Table 3 lists the modal grain size of all quartzites, siltstones, and detrital quartz (where it occurs as an accessory mineral in carbonate and chert rocks), as visually estimated from the thin section. In the Quadrant, the estimated modal values of all samples fall in the sandstone class, with an average for the fifteen values of 0.144 mm. The rocks composed largely of detrital quartz in the Phosphoria formation fall mainly in the silt size class, and have an average estimated modal value for ten samples of 0.042 mm; however, there are a few beds, microveins (in carbonate rocks), and scattered grains which are in the sand size class. The average estimated modal values for the sand beds is 0.134 mm.

The unique frequency curves are, as a whole, smooth and symmetrical. The most noticeable variation is in the peakedness of the curves, with the curves having the better sorting values showing a greater peakedness. The Mo values listed in table 2 range from 0.129 to 0.196 mm. The Mo values in all samples are slightly larger than or equal to the corresponding Md values. The skewness values (Sk, table 3) are all less than or equal to one, ranging from 0.66 to 1.00; these values indicated that the samples are skewed toward the fine fraction.

Heavy Minerals

The heavy minerals of the Quadrant and Phosphoria formations are few in number and type; they comprise from

TABLE 3.—*Modal grain size of quartz grains as observed directly in thin sections*

Detrital quartz constitutes the majority of the rock sample except where the percentage is noted

Sample	mm	Sample	mm	Sample	mm
		Phosphoria			
				(26%) K22a	.30
				(20%) K21A	.17
				K18	.06 st
				K17	.19 qz
			qz		
		S15	.14 (vein)	K16b	.04 st
		S15	.05 st	K16a	.04 st
		S12	.12 qz	K14	.02 st
		(8%) S11a	.04 c	K9a	.03 st
		S10c	.04 st	K8	.12 qz
(8%) W20	.13 i	S8b	.015 st	K7	.12 qz
(Tr.) W16h	.03 i	Sb6	.05 st	K6	.11 qz
(1%) W16e	.94 i	S4	.25* st	K5b	.03 st
W13a	.16 qz	S2a	.08	K4a	.11 qz
		Quadrant			
W12g	.16 qz	S1d	.16 qz	K3b	.12 qz
W12e	.14 qz	S1a	.16 qz	K1d	.10 qz
W12c	.18 qz			K1a	.10 qz
W12a	.20 qz				
W8	.12 qz				
W7a	.13 qz				
W5	.16 qz				
W4	.13 qz				
W3a	.12 qz				
W1	.18				

* Lath quartz.
qz—quartzite.
st—siltstone.
i—insoluble detritals in carbonate rocks.
c—detritals in chert.

0.05 to 0.10 percent of the total weight of the rock. The percentage is usually higher in the Phosphoria where there is an abundance of authigenic collophane. Thirteen varieties of minerals are present: three opaque minerals (hematite, limonite, and leucoxene) and ten nonopaque minerals (anatase, apatite, collophane, garnet, glauconite, muscovite, rutile, titanite, tourmaline, and zircon). The opaques comprise from 20 to 50 percent of the total number of grains; zircon and tourmaline make up 98 percent of the nonopaques. All mineral types, except titanite, are present in both formations with minor variations in morphology and variety. The heavy minerals (sp gr. greater than 2.90) from 15 specimens were separated from the light minerals and examined; detailed counts were made of nine of the heavy separates.

Mineral Suites

The various types of tourmaline and rarer heavy minerals can be grouped into four suites. The classification of the tourmalines is based on that of Krynine (1946).

1. The first suite consists of subrounded to well-rounded grains of tourmaline, zircon, and rutile, which were derived from older sediments. The tourmalines are well-rounded and range in size from 0.04 to 0.15 mm in size. Occasionally a rounded grain is found which has a colorless authigenic overgrowth in optical continuity with the detrital tourmaline grain. The over-

growths are always rounded—evidence that they originated in a previous sedimentary cycle. All 17 varieties discussed below are found in this group. Rounded zircons, usually colorless, are grouped in this suite. The zircons range in size from 0.03 to 0.15 mm. Rounded yellow, orange, and red rutile grains form the third, numerically least significant, member of this group.

2. The second suite consists of those minerals that probably came from an area of high-rank metamorphic rocks, including contact aureole and pegmatitic minerals. Colorless, angular garnets, ovoid appatite grains, and angular to subangular xenomorphic fragments of blue and mauve colored tourmalines are found in this assemblage. Purple zircons, which are considered by British petrographers (Mackie 1923; Boswell, 1927; Brammall, 1923) to have been derived in Great Britain from Pre-Cambrian metamorphics, are also placed in this suite.

3. The third suite is composed of those minerals that are considered to have come from low-rank metamorphic rocks. Round, oval shaped muscovite flakes, small, angular, xenomorphic grains of brown, colorless, and olive tourmaline and very small hypidiomophic grains of colorless tourmaline are placed in this group. Both types of tourmaline frequently contain carbonaceous inclusions.

4. The fourth suite consists of the authigenic minerals. Large angular collophane grains are the most abundant. Apatite, anatase, and glauconite are common; grains of yellow authigenic rutile are rare.

The first suite, consisting of minerals derived from previous sediments, is by far the most abundant; the authigenic assemblage is the second most common; the low-rank metamorphic assemblage is the third most abundant; the high-rank metamorphic assemblage is relatively rare, except for the purple zircons, which are more stable than most of the heavy minerals from this type of rock. (It is by no means certain that these zircons had a metamorphic origin.) The low rank metamorphic and the authigenic suites are slightly more abundant in the Phosphoria than in the Quadrant.

Detrital and Authigenic Minerals

Anatase.—Anatase is present in both formations as authigenic outgrowths on detrital leucoxene. Crystals are pale yellow, well-developed, tabular, and commonly have their corners truncated by the face. Fine-grained, dark inclusions give many of the crystals a dirty appearance. Striated crystals are rare.

Apatite.—Both authigenic and detrital apatite types are present. The detrital grains are well-rounded, oval shaped grains that commonly occur in the less than 0.044 mm size. The grains are colorless, free of inclusions, and have the typical low birefringence of apatite. The authigenic grains are colorless, angular, have an irregular prismatic shape, and contain oval shaped bubble inclusions one to five microns in length. These bubbles generally take up 40 to 60 percent of the surface area of the grain.

The two types are present in both the Quadrant and Phosphoria formations. The detrital grains are more abundant and the authigenic grains are larger.

Collophane.—Authigenic collophane is concentrated in the coarser grain sizes. The grains range in size from 0.02 to 1.0 mm. The irregularly shaped grains are angular and vary in color from nearly colorless through various shades of tan and rust to dark-brown. Collophane has been described in detail under major constituents.

Garnet.—A few grains of garnet are found in samples S1a, S15, and K17. The grains are colorless, subangular, and have the typical, hackly fracture of garnet.

Glauconite.—Glauconite is rare but is present in both the Quadrant and Phosphoria formations. It is most abundant above the lower phosphate horizon. It can be divided into two main varieties: (1) Subspherical, grass-green grains which have distinct, rounded borders. This type averages 0.06 to 0.08 mm in

diameter. The grains contain abundant cracks which are occasionally filled with microcrystalline quartz. (2) The second type consists of irregular patches, composed of aggregates of grains, which conform to the boundaries of the surrounding grains. Both types show aggregate polarization under crossed nicols. The second type is found only above the lower phosphate horizon.

Hematite and Limonite.—The iron oxides are abundant in all samples, but generally do not occur as pure grains. They most frequently are found as coatings on quartz grains and as stains on grains of leucoxene.

Leucoxene.—This is one of the most common minerals—making up about 15 to 30 percent of the total heavy minerals by number. The grains are opaque with a dull white luster, rounded to subrounded, and the surface is generally rough ("porous looking"), though smooth grains are common. Many grains have a yellowish-brown color caused by iron oxide stain. Authigenic, idiomorphic anatase outgrowths are frequently found on the detrital leucoxene grains.

Muscovite.—Muscovite occurs in all samples as oval-shaped, well-rounded flakes averaging 0.08 to 0.1 mm in the direction of the longer diameter. The thin flakes show a light-gray interference color with a rim one to five microns wide around all flakes, showing a white to yellow interference color. Some peculiar set of conditions was evidently in operation during both Quadrant and Phosphoria time to round such supposedly easily floated mica flakes—such conditions are not too common in geologic history. According to Krynine (1937) well-rounded micas indicate "a specialized set of sluggishly moving currents with a gentle to-and-fro motion."

Rutile.—Rutile forms an average of five percent of the nonopaque heavy minerals in both formations, ranging from one to nine percent. It occurs in three distinct colors: foxy-red, orange, and golden yellow. The three types generally occur in approximately equal amounts in each sample.

Titanite.—Titanite is a very rare constituent of the Quadrant heavy minerals. It is easily identified by its high birefringence and characteristic diamond shape. The grains are 0.02 mm long and pale yellow in color.

Tourmaline

Tourmaline is the most abundant heavy mineral in the Quadrant and the second most abundant in the Phosphoria. The wide variation within the tourmaline family makes this an especially easy mineral to classify into types. The heavy minerals are divided into three size ranges: (1) 0.149 to 0.074 mm, (2) 0.074 to 0.044 mm, and (3) less than 0.044 mm. The tourmalines are most abundant in the 0.074 to 0.044 mm size range. The tourmalines can be further subdivided on the basis of the grain shape: (1) rounded (no angular corners), (2) subrounded (one third of the corners angular) to subangular (two thirds of corners angular), and (3) angular (all corners angular). A final subdivision can be made on the basis of color and inclusions.

Round Varieties.—The rounded types make up an average of 13 percent of the Quadrant tourmalines and nine percent of the Phosphoria tourmalines in the 0.074 to 0.044 mm size range. The extreme rounding in this group and the presence of abraded authigenic overgrowths indicates that they were derived from reworked sediments. The rounded grains are concentrated in the two coarser grain sizes, being completely absent in the pan fraction. There is a pronounced increase of rounded grains in the coarser grain sizes.

Angular Varieties.—The angular types comprise an average of 15 percent of the Quadrant and four percent of the Phosphoria tourmalines in the 0.074 to 0.044 mm size range. Angular grains are absent in the coarser size and increase in abundance in the finer sizes. Most of the tourmalines in this group are xenomorphic,

though some are hypidiomorphic. The hypidiomorphic grains are always colorless.

Subangular to Subrounded Varieties.—This group contains the great bulk of tourmaline grains; 72 percent of the Quadrant and 87 percent of the Phosphoria tourmalines in the 0.074 to 0.044 mm size range are in this group. This group is slightly more abundant in the coarser grain sizes.

The tourmalines can be subdivided into 17 varieties on the basis of color and inclusions. All 17 varieties of tourmaline are found in each of the three shape classes. The fact that all 17 varieties occur in each grain shape group and the fact that there is an increase in the number of rounded grains in the coarser size and an increase in the angular grains in the finer grain size suggests the possibility that this variation in grain shape is due to selective rounding rather than differences in source area.

Tourmaline Varieties

The tourmalines can be divided into 11 color types:

1. Brown tourmaline.
2. Chocolate-brown tourmaline.
3. Green tourmaline
4. Black tourmaline—nonpleochroic basal sections that show good interference figures.
5. Near colorless tourmaline—pleochroic from colorless to light tints of yellow, tan, pink, gray, green, or blue.
6. Multicolored tourmaline—grains largely brown with large blue, or occasionally colorless, spots.
7. Blue tourmaline.
8. Mauve tourmaline.
9. Olive tourmaline.
10. Gray tourmaline
11. Chestnut-brown tourmaline.

The tourmalines described above were classed into 17 easily distinguishable varieties based on color and inclusions. Detailed grain counts were made in the 0.074 to 0.044 mm size range.

The number of tourmaline grains counted ranged from 125 to 475. There is no consistent, significant difference in the percentages of the 17 various tourmaline types within either the Quadrant or Phosphoria formations, and very little between formations. There is a difference in the relative abundance of near colorless tourmaline containing a few inclusions between the two formations. The four samples of the Quadrant contain an average of 14 percent of this type of tourmaline, and the five samples in the Phosphoria contain an average of 28 percent. To test the significance of these values, it was necessary to compare the spreads or variabilities of the two sets of figures. The Fisher F test was used to do this. A probability value of 0.04 was obtained, which means that only four times in 100 could a difference that large occur due to chance.

The star diagrams in figure 11 are based on a plot of the six basic color subdivisions of the tourmalines from both formations. The diagrams are quite similar, the largest difference being the increase in colorless tourmalines in the Phosphoria.

Zircon and Its Varieties

Zircon is the most abundant nonopaque heavy mineral in the Phosphoria and the second most abundant in the Quadrant. Zircon is concentrated in the two finer size ranges; the 0.074 to 0.044 mm size range was used for grain counts. The zircons can be subdivided on the basis of degree of rounding: (1) rounded, (2) subrounded, and (3) subangular. The grains were next divided on basis of morphology: (1) equant and (2) prismatic. Other divisions can be made on the basis of color: (1) colorless and (2) purple-pink, or inclusions; and finally surface features: (1) pitted and (2) smooth. The number of grains counted ranged from 136 to 435 except for sample W7a from which only 30 grains were counted.

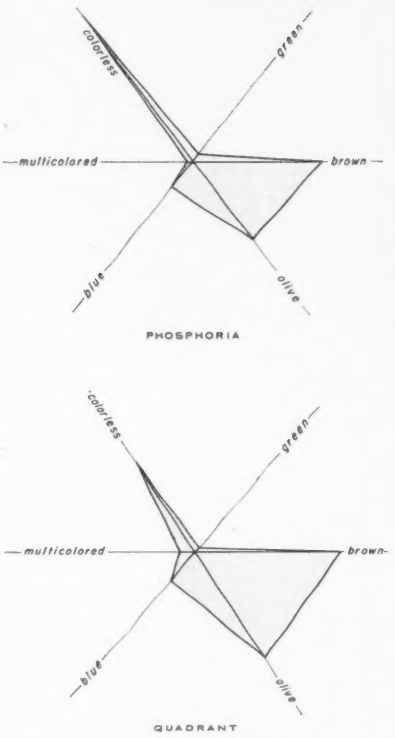

FIG. 11.—Star diagram showing the relative distribution of the six major color types of tourmalines.

There is a greater percentage of rounded zircons in the Quadrant than in the Phosphoria, and conversely, a greater percentage of subrounded grains in the Phosphoria than in the Quadrant. A probability or significant value of 0.05 was obtained for the variation in the number of subrounded grains between the two formations. The grains are about 60 percent prismatic and 40 percent equant in both formations. All grains show some rounding. There is a tendency towards idiomorphism in some grains in the pan (less than 0.044 mm) size.

Colorless zircons comprise 91 percent of the total zircons in both formations. Smooth grains are scarce, the majority showing some pitting. Inclusions occur in many of the grains, the inclusions are generally bubbles; zoned grains are not common. The most significant difference between the two formations is in the colorless variety with inclusions, which constitutes 45 percent of the zircons in the Phosphoria and only 30 percent of the Quadrant zircons. (Colorless tourmalines, with inclusions, are also more abundant in the Phosphoria than in the Quadrant.) The significant value for this difference is only 0.20.

The purple-pink variety of zircons forms nine percent of the total zircons in samples from both formations, ranging from six to 13 percent. The color varies from a light tint to a deep striking color. Fresh, unpitted grains are rare; inclusions, similar to those in the colorless variety, occur in approximately half of the grains.

Yellow zircons are present in many samples, but are rare.

Zircon-Tourmaline Relations

There is a difference in the ratio of tourmaline to zircons in the 0.074–0.044 mm size range of the two formations.

Number of Zircons Counted	Tourmaline / Zircons Ratio		Percent of Zircons
237	K4a	—0.18	85
213	K8	—0.20	83
316	K17	—0.35	74
175	S6b	—0.30	71
230 Phosphoria	S15	—0.30	77
167 Quadrant	S1a	—0.86	54
435	W3a	—1.1	48
30	W7a	—6.6	13
136	W12d	—1.0	49

This ratio holds true in six other samples examined but not counted in detail. The percentage of zircons in the less than 0.044 mm size is greater than in

the coarser sizes (in both formations); however, the difference in the relative amounts of zircons between formations is the same. There are no major differences in the Md values of the sands from the two formations; however, the sands from the Phosphoria have approximately twice as much material in the silt size as the Quadrant sands; this may be the reason for the greater percentage of zircons in the Phosphoria, as the amount of zircons generally increases as the enclosing sand size decreases.

One of the most common heavy minerals in altered volcanic material is zircon. Other evidence has suggested that there was an abundance of volcanic material in the Phosphoria source area. It is possible that this is, in part, the reason for the increased zircon content in the Phosphoria.

Rock Types

General

The Quadrant sand is a blanket-like sand of wide lateral extent, approximately 70,000 square miles (including the closely associated Tensleep), and moderate thickness, averaging 200 to 500 feet. In the area described, the Quadrant is a uniform detrital quartz sand; it is a massive sand with some cross-bedding and only a trace of graded bedding on a microscopic scale. The heavy minerals are, as a whole, typical of well-reworked sands. The Quadrant sand can be classified as an orthoquartzite.

The Phosphoria formation is also a blanket-like body with a lateral extent of approximately 175,000 square miles and a thickness ranging largely between 100 and 400 feet. The Phosphoria is much more variable in its makeup than is the Quadrant and it is characterized in this area by an abundance of chemical rocks.

Several individual beds, a few feet thick, of sand size detrital quartz rocks occur in the Phosphoria at Kelly Gulch. These rocks are similar to the orthoquartzites of the Quadrant except for the increased amounts of cements or matrices (microcrystalline quartz, dolomite, and illite). These quartzites as a whole are slightly finer grained and less well-sorted than the Quadrant quartzites. These orthoquartzites can be classed as:

1. Dolomitic orthoquartzite
2. Silicious orthoquartzite
3. Illitic orthoquartzite

Siltstones are the most abundant rock type in the lower Phosphoria. At Wadham's Spring there are only a few beds although at the other two trenches siltstones are dominant. The siltstones belong to the same types as the orthoquartzites; however, whereas illite occurs as a matrix in the siltstones at Kelly Gulch, kaolinite is the principal clay matrix at Sheep Creek and Wadham's Spring. Dolomitic siltstone is most abundant at Kelly Gulch.

The siltstones generally show good bedding ranging from 0.5 to six inches wide. Thin, one mm laminae are frequently present, outlined by various colored iron stains. The siltstones commonly contain considerable limonite stains and a few iron nodules (five to 10 mm in diameter).

Carbonate rocks are the second most abundant rocks in the Phosphoria and, of these, dolomites and dolomitic limestones comprise the bulk with pure limestones occurring only as thin, hard, irregular, undolomitized beds within the dolomitized rocks. Carbonate rocks are the predominating rock type at Wadham's Spring; they also occur in abundance at Sheep Creek, but at Kelly Gulch carbonate minerals are present only as a cement. The carbonate rocks commonly have a one to four inch bedding.

Chert, occurring both as nodules and beds, has been described in detail. Nodules are found in all of the carbonate rocks; bedded chert is most abundant in the Sheep Creek trench.

Oolitic phosphate rock, in thin beds one to two inches thick, occurs 35 to 125 feet above the Quadrant-Phosphoria contact. The phosphate rock is intimately associated with bedded chert.

Quadrant Quartzite

Fine-Crained Orthoquartzites.—Nearly all of the Quadrant samples have an Md value that falls in the fine sand range and generally 50 percent or more of the grains are concentrated in this narrow range with most of the remaining grains falling in the finer sizes. The sand is extremely well-sorted and moderately rounded, appearing as a typical orthoquartzite in thin sections.

The average orthoquartzite contains 93 percent detrital quartz, four percent detrital chert, two percent quartz overgrowth, and one percent clay—there is very little variation from this average composition. The clay is largely kaolinite with a few shreds of illite. The quartz overgrowths occur on the majority of the grains but are only moderately well-developed. The sand grains are held together commonly by the quartz overgrowths, but frequently tight packing alone holds the grains together. The Quadrant orthoquartzite is only a moderately hard rock, except at Kelly Gulch where minor metamorphism has welded the grains together to such an extent that the rock approaches in texture a metaquartzite.

Phosphoria Formation

Fine- to Very-Fine Grained Orthoquartzites.—The Md values of the few Phosphoria quartzite samples generally fall near the dividing size of fine and very fine sand (0.125 mm) with about 70 percent of the grains being equally divided between the two sizes. The majority of the remaining grains falls into the smaller size groups. The sands are well-sorted but not as well-sorted as the Quadrant. The majority of the grains fall into the subround to rounded range. In the orthoquartzites detrital quartz grains average 78 percent and detrital chert four percent; the grains are held together by a variety of cementing minerals.

1. Dolomitic orthoquartzite.—The dolomite cement is present in amounts ranging from 15 to 20 percent. Some of the dolomite grains grow together in a true cement fashion, but many of them occur as isolated grains and stringers, seeming more like a matrix than a cement. Some of the dolomite may be clastic. About 0.5 percent illite, occurring as individual shreds and aggregates and a few kaolinite books are commonly present. A few flakes of altered biotite and a few detrital, rounded, fine-grained calcite grains are materials peculiar to this type of rock.

2. Siliceous orthoquartzite.—The siliceous cement is composed of microcrystalline quartz (chert) which comprises from 15 to 35 percent of the rock. The chert acts as a true cement, completely surrounding the detrital grains. Generally one to two percent of irregular aggregates of kaolinite are present. One sample contains 10 percent hematite and limonite associated with the chert cement.

3. Illitic orthoquartzite.—This name is applied to detrital quartz rich rocks that contain a considerable amount of authigenic illite. At Kelly Gulch illite makes up 10 to 15 percent of the quartzite rock. The illite occurs as individual shreds and narrow bands between grains, and as books and irregular patches up to 0.080 mm wide. The illite sometimes replaces portions of quartz grains and the associated dolomite cement.

4. Orthoquartzite.—One orthoquartzite at Kelly Gulch contains no matrix and is similar to the Quadrant orthoquartzite. There is a trace of quartz overgrowths, but most of the bonding is due to minor welding of the slightly crenulated borders.

Coarse Siltstone.—The Md value of the siltstones range from 0.15 to 0.06 mm, averaging 0.04 mm. Frequently one or two percent of sand size grains with a maximum size of 0.24 mm are present. The siltstones are only moderately well sorted and the majority of the grains are either subangular or subrounded. The siltstones differ very little from the quartzites described above. The detrital

quartz in the siltstones averages 80 percent; detrital chert is present only in traces.

The amount of dolomite present in the dolomitic siltstones averages 23 percent with little variation between samples. One to five percent of illite or kaolinite occurs with the dolomite. One sample, K14, has only illite as a matrix—20 percent. At Sheep Creek kaolinite acts as a matrix; it is present in amounts ranging from three to 30 percent, averaging 15 percent. The kaolinite contains a few shreds of illite and occurs in the same manner—kaolinite books are conspicuous.

Limonite stains and muscovite shreds are generally found in the siltstones.

Carbonate Rocks.—1. Limestones. Hard, dense, medium-brown beds of limestone generally less than a foot wide occur regularly in the dolomitized rocks and show an irregular rough contact with these. The moderately sorted calcite grains average 0.4 mm in diameter. The limestones contain approximately 10 percent insoluble matter which is generally detrital quartz, silt size with some kaolinite and considerable authigenic chert. Pure limestones are found only at Wadham's Spring.

2. Dolomitic limestone. These rocks are abundant at Wadham's Spring. They are medium-hard, tan beds a few feet thick. The dolomite is present as hypidiomorphic dolomite grains averaging 15 to 30 microns in diameter, in a matrix of coarse-grained, xenomorphic calcite, averaging 0.4 mm. The dolomitic content ranges from 10 to 60 percent, generally being near 50 percent. The insoluble residue again averages 10 percent, ranging from two to 20 percent. The residue is the same as in the pure limestones except for sample W19b where montmorillonite is present. Dolomite breccias are common in this rock type.

3. Dolomite. Dolomite is abundant both at Wadham's Spring and Sheep Creek. The dolomite varies from soft- to medium-hard and has a dull-white to light-tan color. It is well-sorted, with a grain average between 10 and 20 microns. The insoluble residue is similar to that in the limestone and averages eight percent for all samples except two which contain 30 to 35 percent.

Chert.—The chert rocks generally contain abundant replaced fossils, a few grains of detrital quartz, one or two percent carbonate grains, limonite stains, and occasionally shreds of illite. The cherts near the phosphate rock commonly contain grains of collophane and glauconite.

Phosphate Rock.—Phosphate rocks vary from white to green to black in color. Their collophane content is quite variable, changing from five to 90 percent within a few millimeters. Collophane generally occurs as oolites, bone fragments, and cement. Detrital quartz and chert grains within the phosphate rocks are of medium sand size. Authigenic chert is intimately associated with the collophane; usually the two interfinger.

Stratigraphic Sections

Figure 2 shows the three stratigraphic sections of the area. The sections, based on field notes and considerable laboratory work, are believed to be fairly accurate. Boundaries between formations were determined by members of the U. S. Geological Survey.

At Wadham's Spring the boundary between the Quadrant and Phosphoria formations is placed at the first appearance of chert nodules and irregular chert fragments, although the host rock still is identical with the Quadrant quartzite.

At Sheep Creek the contact is placed at a definite change in lithology from Quadrant quartzite to dolomite and bedded chert.

At Kelly Gulch the Quadrant quartzite was separated from the Phosphoria on the basis of a lack of resistance to weathering of the Phosphoria. This is a valid distinction as the quartzite above the boundary contains an abundance of

TABLE 4.—*Percentage of rock types comprising formations and areas*

	Quadrant Average	Phosphoria Average	Wadham's Spring	Sheep Creek	Kelly Gulch
Quartzite	100%	11%	10%	0	22%
Siltstone	0%	52%	.05%	80%	75%
Carbonate	0%	35%	90%	14%	0%
Bedded Chert	0%	2%	0%	5%	2%
Phosphate Rock	0%	1%			
Sand/silt Ratio	100:1	1:5			

illite, five to 20 percent, and dolomite, 20 percent, which is missing from the quartzite below the boundary. The quartzites, aside from this illite and dolomite, are identical.

If the oolitic phosphate beds in all three trenches were deposited at the same time, the surface of the Quadrant orthoquartzite would be approximately 75 feet higher at the Sheep Creek area (Middle section) than at either of the other two areas. This high area would form an effective barrier separating two shallow basins. Such an arrangement would help to account for the lithologic differences in the Kelly Gulch and Wadham's Spring areas.

PETROLOGY AND ORIGIN OF THE QUADRANT AND PHOSPHORIA FORMATIONS

General

In comparing and reconstructing the origin of two sedimentary formations, it is necessary to examine, primarily, variations in the source areas and in the methods of deposition. In order to do this, it is necessary to infer from the properties of the sediments the physical and/or chemical processes that are responsible for the sedimentary parameters and to evaluate whether this is a reflection of conditions in the source area or in the method of deposition—or a combination of the two.

The broadest characteristic of a sedimentary formation is its shape. In this case, both formations are blanket-like which implies that there was no extreme tectonic change during the deposition of the two formations. However, in another major characteristic, lithology, there is a definite difference.

The ratio of detrital to authigenic material is 98:2 in the Quadrant, and 48:52 in the Phosphoria (only the material below the phosphate beds is used in the Phosphoria computations). The percentage of the rock types comprising the formations and individual areas are shown in table 4.

There is a decided increase in chemical rocks and an even greater increase in total chemical material in the Phosphoria as compared to the Quadrant. However, within the Phosphoria there is an even greater difference. Ninety-seven percent of the Kelly Gulch rocks are quartzites and siltstones whereas 90 percent of the Wadham's Spring rocks are carbonates. The Sheep Creek rocks type distribution is intermediate between these two. The abundant sand and silt detritus at Kelly Gulch likely represents a near-shore environment, and the carbonate rich Wadham's Spring area represents a more seaward area of deposition. The lithologic differences in these two areas was perhaps enhanced by the Sheep Creek ridge which would have prevented most of the detrital material from reaching the Wadham's Spring area.

In the quartzite rocks of both formations, the types of detrital quartz show little variation; there is slightly more end phase and metamorphic quartz in the Quadrant. Plutonic quartz was separated into three varieties on the basis of types

of bubble inclusions; these varieties occur in approximately the same ratio in both formations. Microlites were of the same type and the same relative abundance in both formations. Volcanic quartz occurs only in the Phosphoria.

Detrital chert was divided into three types on the basis of the size of the microcrystalline quartz. The relative amounts of each type were the same in both formations.

The heavy minerals are usually a sensitive indication of changes in the source area. In the heavy minerals from the two formations, the ratio of colorless to purple zircons are the same in both formations. Within the colorless class there is an increase in the percentage of colorless zircons with inclusions in the Phosphoria. In the 17 types of tourmalines, one type, colorless with inclusions, is more abundant in the Phosphoria. The zircons are less well-rounded and more abundant in the Phosphoria. The proportion of prismatic to equant zircons is the same in both formations.

Five other detrital heavy minerals, occurring in trace amounts, are present in both formations. A few grains of biotite and microcline are present in the Phosphoria and not in the Quadrant.

It seems likely from this that the source area for the detritus of the Phosphoria was much the same as that for the Quadrant. However, it is likely that the Phosphoria source area contained some Quadrant material. The abundant silty detritus in the Phosphoria is interfingered with sand size detritus similar to the Quadrant sand. This suggests a general lowering of relief in conjunction with a fluctuating shoreline.

Nothing was found in the Quadrant that indicated the direction of the source area. However, in the Phosphoria the increase in the amount of quartzites from south to north and the increase in chemical rocks and chemical material from north to south suggests that the source area lay somewhere to the north.

The conditions of deposition are best inferred from the degree of rounding and sorting of the sand. The Quadrant quartzites are slightly better rounded than the Phosphoria quartzites. The quartzites in both formations fall in the well-sorted beach sand class, although the Quadrant sand is better sorted than the Phosphoria.

The abundant thin-bedded, non-cross-bedded, siltstones in the Phosphoria indicate that there was little reworking of detritus on the sea floor. However, the presence of reworked carbonate (Wadham's Spring) formed within the basin of deposition shows that there was a mild reworking of some of the Phosphoria material.

The few fusulinids occurring in the Phosphoria suggest that the water was relatively shallow (on the order of 180 feet as interpreted by Elias (1937)) in the basin of deposition.

The chemical rocks formed within the basin of deposition directly reflect the conditions within the basin and indirectly reflect the conditions of the surrounding source area. The abundance of these chemical rocks in the Phosphoria indicates low relief and relative quiescence.

Evidence indicates that calcite was first deposited, microcrystalline quartz began to replace the calcite, but before this proceeded far, dolomite replaced much of the remaining calcite and was in turn replaced, in part, by the microcrystalline quartz. Collophane later replaced some of the microcrystalline quartz. Some of the completely dolomitized material was reworked on the sea floor and incorporated in later, only partially dolomitized, limestone.

The origin of the clay minerals is difficult to deduce. They are present in 95 percent of the rocks in the Phosphoria and commonly in amounts as high as 30 percent. Many of the clay minerals occur as large books, worms, and oriented "nests"; They are also found replacing carbonate minerals and quartz grains. This mode of occurrence and the lack of

any shale rocks in this portion of the Phosphoria strongly suggest that much of the clay is authigenic.

Kaolinite is greatly predominant over illite in both the Wadham's Spring and Sheep Creek areas. However, it is much more abundant at the Sheep Creek area. Illite predominates over kaolinite at the Kelly Gulch area.

In a typical quartzite-limestone sequence, clay minerals are usually relatively scarce, particularly when there is no evidence of metamorphic or feldspathic rocks in the source area. There is a source material for clay minerals that is usually not reflected in the coarser sedimentary detritus—volcanic lavas and tuffs. Volcanic ash could be deposited directly into the sea and be altered to form clay minerals, but more likely most of the ash and lavas would be deposited in the surrounding peneplained source area. In a highly peneplained region where chemical weathering is intense, the volcanic material would be taken into solution and the alkalis, alkaline earths, silica, alumina, and iron carried to the sea where they would recombine to form clay minerals. The type of clay mineral formed would vary locally with the variation in environment in the basin of deposition.

Kaolinite is found in an acid environment and is most commonly found in continental sediments; it is not believed to form under marine conditions. Illite is formed in a basic environment and is a common marine clay. However, if sufficiently acid waters could occur in a marine environment, there is no reason why kaolinite could not form. Such a condition may occur on the top of topographic highs on the sea floor. Holzman (1952), in his study of the Cortes and Tanner Banks off the Coast of Southern California, found that the water on top of these banks was sufficiently acidic, pH 6.6 to 6.9, to dissolve tests of foraminifera. Emery and Rittenberg (1952) in their study of a dozen basins off the Southern California coast reported the pH values ranging from 7.3 to 8.5 in the basins. It seems possible that a shallow water acid environment on top of the Sheep Creek ridge could cause the formation of kaolinite, whereas in the relatively normal, basic environment in the Kelly Gulch basin illite would form.

Though there is little direct evidence to indicate abundant volcanic material in the source area, its presence can be used to explain nearly all the anomalies in the Phosphoria formation. Mansfield (1927, 1940) told of abundant Permian, rhyolitic to andesitic, volcanic material in eastern Oregon and west and central Idaho, and suggested that this was the source of the fluorine in the phosphate minerals and the silla in the Rex chert. Mansfield considered that most of the fluorine came from volcanic gases and that the silica came from weathering of volcanic tuffs and lavas.

The detrital and carbonate material in the Phosphoria formation is typical of a normal second-cycle quartzite-limestone series, in which the source material is older quartzites and limestones. The material anomalous to such a tectonic and environmental background is authigenic clay minerals, abundant chert, concentrations of phosphate, vanadium, and nearly all the trace elements.

Clarke's (1924) analyses of rhyolites and andesites, particularly the andesites, show a higher concentration of nearly all trace elements than is found in the average sediment. Intense weathering of volcanic tuffs and lavas would provide abundant phosphorous, vanadium, and trace elements to form the concentrations in the Phosphoria. Silica, aluminum, magnesium, iron, and potassium would be carried to the sea and form the large amount of clay material present. The excess silica would be deposited as chert. If a large portion of the source area was composed of volcanic material this would explain the sparsity of detrital material in the Phosphoria.

Reconstructed Picture of Sedimentation on the Northeast Flank of the Phosphoria Sea

The rounded quartz grains, the abundance of detrital chert, and the predominance of the rounded, stable minerals, zircon, rutile, and tourmaline, and good sorting of the sand with little evidence for much reworking within the basin of deposition indicates the source area, to the north, was at least, in part, similar for both the Upper Quadrant and the Lower Phosphoria formations and was composed largely of sediments—quartzites and limestones from early Paleozoic and/or pre-Cambrian Belt rocks.

A small percentage of metamorphic quartz and end phase quartz plus the presence of garnet, apatite, and certain types of angular tourmaline indicates that there were a few metamorphic rocks, both high- and low-rank, and possibly some pegmatitic rocks in the immediate source area.

There was probably some slight emergence and erosion of the Quadrant formation followed by Permian volcanic activity to the west and northwest which covered many of the sediments in the source area with volcanic tuffs and lavas (Eardley, 1951).

The rocks, described excepting the authigenic clays and phosphates, comprise a typical quartzite-limestone series which is characteristic of a stable, fairly well peneplained platform region, containing shallow basins of deposition. There is no sharp break between the sediments of the two formations. The detritals are nearly identical, but there is an increase in the amount of chemical rocks deposited in the Phosphoria, and a general lowering of the surrounding land area as silt became the major detrital size instead of sand.

The fine sand of the Quadrant was presumable transported by rivers having a very low gradient. There was little reworking by wave action; however, the sorting and degree of rounding indicates, assuming similar source areas for the detrital material of both formations, that there was more reworking of the Quadrant than of the Phosphoria sand. The mutual contact of quartz overgrowths show that packing had not proceeded very far when the silica was precipitated.

As the landmass to the north was gradually lowered either by erosion or minor subsidence or both, silt size detritus began to predominate over the fine sand size detritus. Meanwhile, volcanic material to the west and northwest was being intensely weathered and a wide variety of elements carried on into the sea.

At Kelly Gulch, which was close to the northern source area, silt was deposited in abundance while carbonate and illite minerals, forming continuously within the basin, were deposited in amounts of 20 to 25 percent and acted as matrix or cement.

Farther to the south at Sheep Creek and Wadham's Spring and farther from the land, chemical activity became predominant and carbonate minerals increased in abundance. Calcite was deposited first and while still in the form of an ooze was replaced by dolomite and microcrystalline quartz. Soon after this, when the chert and dolomite were still in contact with the sea water, some reworking occurred. The presence of shallow marine fossils in the two southern sections indicates a shallow, oxygenated sea.

As deposition continued, there were minor periodic transgressions and regressions of the sea and thin beds of sand size detritus were deposited within the silts at Kelly Gulch. At Sheep Creek silt size detritus soon became predominant, and chemical activity diminished with only authigenic kaolinite being deposited. Meanwhile, at Wadham's Spring chemical deposition continued indisturbed.

As deposition continued, glauconite became more abundant; collophane bone fragments appeared; phosphate was re-

moved from the sea and deposited as collophane. The collophane was deposited as cement and oolites, and sometimes replaced chert.

In summary it can be seen that the detritus of the two sedimentary formations was primarily derived by the reworking of older sediments. The detrital and carbonate minerals of the two formations comprise a quartzite-limestone series and are typical of a period of near peneplaination in the source area and belongs in the shallow water marine facies of sedimentation. It is indicative of epeirogenic tectonic control both in the source area and in the locus of deposition (Krynine 1948). The rocks in the Phosphoria are modified by the deposition of an abundance of authigenic minerals and a general decrease in detrital material. This modification was likely caused by an abundance of volcanic material in the source area.

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FIGS. 1–4, SEPTEMBER, 1955

CLAY MINERALS OF THE NEUSE RIVER ESTUARY

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ABSTRACT

The clay fractions of 24 Neuse River Estuary bottom samples were analyzed by X-ray diffraction techniques to determine the clay mineral content and to note any diagenetic changes in the clays as they were introduced into a brackish-water environment. Kaolinite and "chlorite" were present in all 24 samples and illite in 18. Kaolinite was by far the dominant clay mineral being introduced into the estuary, but it decreased considerably relative to "chlorite" and illite as increasingly saltier water was encountered downstream. "Chlorite" increased downstream becoming the dominant mineral at the lower end of the estuary. Illite, though very subordinate to both kaolinite and "chlorite" throughout the estuary, showed a definite increase in the last few downstream samples and appeared to be increasing more rapidly than "chlorite."

The "chlorite" was thermally unstable and collapsed to a 10 Å structure when heated at 400° C for one hour; however, it was like normal chlorite in retaining the 14 Å basal spacing when treated with ammonium salts or glycerol.

INTRODUCTION

This paper is a report on the clay mineral content and the diagenetic changes in the clays of the Neuse River Estuary of North Carolina as determined by X-ray diffraction analyses. This estuary is in the brackish water transition zone between the fresh waters of the Neuse River and the marine waters of the Atlantic Ocean. Here in the estuary are to be found the first adjustments of the fresh water clays to marine conditions. This estuary was chosen for study because of the completeness with which its fresh water clays are known (Brown and Ingram, 1954).

THE NEUSE AND ITS DRAINAGE BASIN

The Neuse River is one of the many moderate sized rivers draining the south Atlantic slope of the United States. The Neuse is about 180 miles long extending from near Durham, North Carolina, southeast to the Pamlico Sound. In the drainage basin is a great variety of rock types that includes Pre-Cambrian(?) gneisses and schists; Ordovician (?) volcanics and metasediments; Carboniferous (?) granites; Triassic conglomerates, sandstones, mudrocks, and diabases; Cretaceous sandstones, mudrocks, and limestones; and Cenozoic gravels, sands, clays, and limestones (fig. 1).

Two miles above New Bern, the Neuse broadens into an estuary (fig. 2). The estuary is about 40 miles in length and gradually increases in width from about one mile at New Bern to about six miles at its entrance into Pamlico Sound. In the main channel the water increases in depth from about 10 feet at the head of the estuary to about 25 feet at the entrance into Pamlico Sound.

Tides are practically nonexistent in the Neuse River Estuary, but easterly winds sometimes cause water level increases of as much as four feet at New Bern (U.S.C. & G.S., 1939). The absence of tides prevents the contamination of the river derived estuarine sediments by marine sediments and increases the probability that all the differences between the river clays and the estuarine clays are caused by diagenesis.

Not enough work has been done to establish a good picture of the salinity and circulation patterns of the estuary, but the limited data available (table 1) show that there is some degree of salinity stratification. In general, the bottom waters are more saline than the surface waters, but occasionally this stratifica-

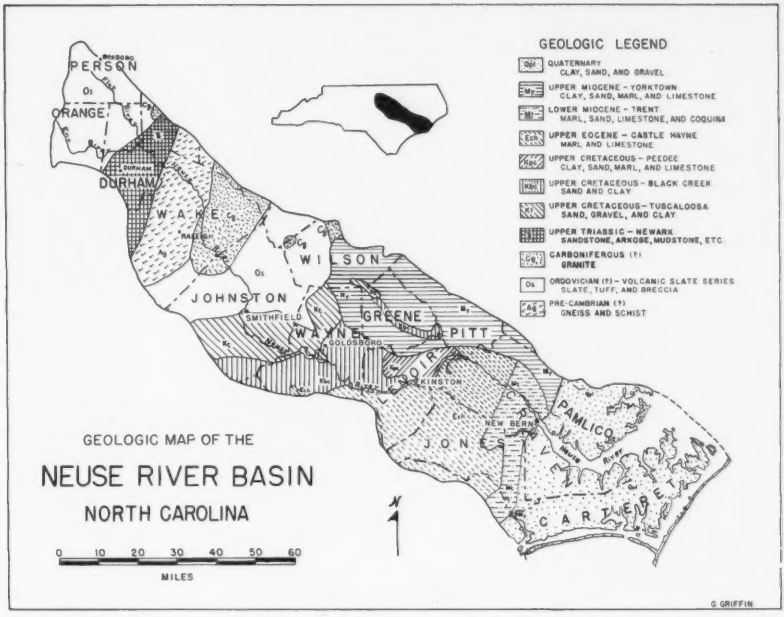

FIG. 1.—Geologic map of the Neuse River drainage basin. Modified from maps by Berry (1947) and Mundorff (Riley, 1950).

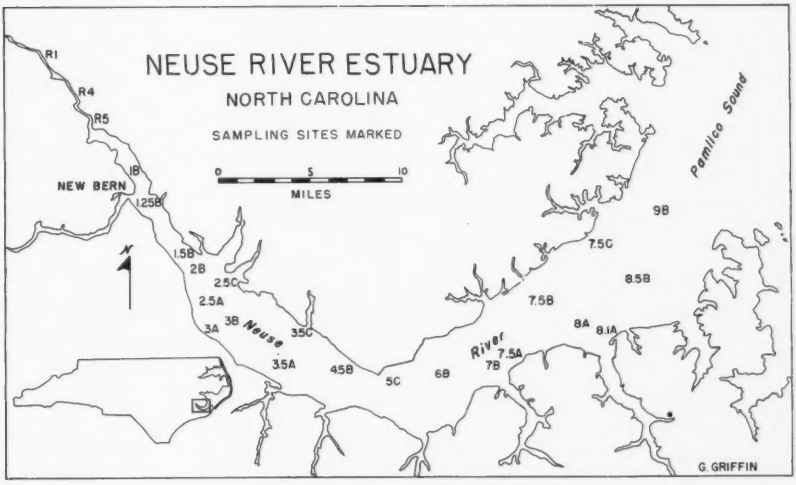

FIG. 2.—Sampling sites in the Neuse River Estuary.

TABLE 1.—*Salinity in the Neuse River Estuary*

Sample Location	Miles below Street's Ferry (R1)	Date (mo/yr)	Salinity °/₀₀ (surf./bottom)	Source of Information
Above New Bern (R5)	5	4/29	0.14*	Riley, 1950, p. 91
New Bern (1B)	8	2/47	0.46*	Lamar and Joyner, 1949, p. 32
Otter Creek Buoy (3B)	16	5/53	4.55*/	Griffin
Garbacon Shoal Buoy (7.5B)	35	6/48	9.92/16.74	Roelofs, 1953
		8/48	18.50/20.10	
		9/48	18.68/20.84	
		12/48	2.41/15.39	
		3/49	5.30/ 6.30	
		6/49	6.93/ 7.14	
		9/49	8.00/ 8.20	
		12/49	9.36/12.23	
		9/50	11.33/12.64	
		11/50	7.97/12.47	
		1/51	12.20/14.47	
Point of Marsh Light (9B)	42	8/48	16.19/16.03	Roelofs, 1953
		9/48	20.45/20.45	
		4/49	9.30/11.20	
		6/49	11.07/11.89	
		9/49	11.45/12.00	
		12/49	11.74/11.69	
		9/50	14.57/13.53	
		11/50	10.84/15.38	
		1/51	16.02/16.00	

* Total dissolved solids.

tion is destroyed by vertical mixing. The less dense, fresh waters flow seaward on the surface becoming more saline downstream because of mixing with the bottom waters. The denser, saline waters move upstream along the bottom becoming less dense by mixing with the surface waters. At the mouth of the estuary where the Neuse enters Pamlico Sound, there is little if any persistent salinity stratification. Salinity in the estuary varies inversely with the rainfall in the Neuse drainage basin with great differences in salinity being found in salinity determinations taken a month apart. In general, the maximum salinity is about one-third to one-half that of normal sea water.

Limited pH determinations (table 2) show that the waters of the estuary have about the same degree of alkalinity (pH 8) as normal sea water although the inflowing fresh water of the Neuse River is slightly acid (pH 6).

METHODS OF INVESTIGATION

During the months of May and June, 1953, 24 bottom samples were collected with a clamshell-type sampler. For convenience in locating sampling sites, the estuary was divided into nine sections, the boundaries of the sections being easily recognizable channel markers. Within each section, samples from near

TABLE 2.—*pH in the Neuse River Estuary*

Sample Location	Miles below Street's Ferry (R1)	Date (mo/yr)	pH (surf./bottom)	Source of Information
New Bern (1B)	8	2/47	6.5	Lamar and Joyner, 1949
Garbacon Shoal Buoy (7.5B)	35	8/48	8.02/7.78	Roelofs, 1953
		9/48	8.12/7.84	
		9/48	7.95/7.98	
Point of Marsh Light (9B)	42	8/48	8.08/8.10	Roelofs, 1953
		9/49	8.10/8.15	

the southern shore were designated with an "A"; samples from the main channel were designated with a "B"; and samples from near the northern shore were designated with a "C". Decimals were used to show the approximate position lengthwise within each section. For example, 7.5 B indicates a sample from the channel near the center of section 7. Fresh water river samples were designated with an "R" and numbered consecutively downstream from Street's Ferry, 10 miles above New Bern. Sampling sites are shown on figure 2.

Each sample was dispersed in 0.01 N sodium hexametaphosphate (Calgon), and the clay fraction separated by centrifuging and flocculation with a saturated calcium chloride solution. All samples were oven-dried at 35° C.

X-ray diffraction patterns were made with a Hayes X-ray diffraction unit equipped with 114.6 mm diameter powder cameras. Samples packed in thin-walled cellulose acetate tubes 0.5 in mm diameter were exposed to copper radiations. The K_b radiations were filtered out with nickel foil placed just in front of the photographic film.

X-ray diffraction patterns were made of each clay fraction. When necessary to complete the identification and characterization of clay mineral components, additional patterns were made after the samples had been subjected to glycerol solvation, cleaned with HCl and HN_4OH, or heated at various temperatures.

RESULTS

The lithology and color of each sample are shown in table 3. Samples from the channel are mainly dark muds (silty clays), and the samples between the channel and the shore are mainly very fine to medium sands.

Kaolinite and "chlorite" were identified in all 24 samples, and illite was identified in 18 samples (fig. 3). Considerable amorphous or poorly crystallized material was present in six of the samples. In general, kaolinite is the most abundant clay mineral; but in the lower part of the estuary, "chlorite" and illite are about as abundant as kaolinite.

The "chlorite" present in these samples is perhaps best described as a "diagenetic chlorite" as its properties are not those of ordinary chlorite. Like ordinary chlorite, this "diagenetic chlorite" has a 14 Å basal spacing that does not expand upon glycerol solvation and does not contract upon boiling in a weak ammonium-salt solution. Unlike ordinary chlorite, this "diagenetic chlorite" is not thermally stable at temperatures up to 700° C. When samples were heated in successive 100° C steps, the 14 Å line faded considerably between 200 and 300° C and was completely gone at

TABLE 3.—*Lithology and color of samples*

Sample No.	Lithology	Color*
R1	Silty clay; much woody material	Mod. yellowish brown
R4	Very coarse to medium sand	Mod. yellowish brown
R5	Clayey coarse sand; mica flakes	Mod. yellowish brown
1B	Silty clay; much woody material	Lt. olive gray
1 25B	Silty clay; much woody material	Dk. yellowish brown
1.5B	Clay; many small, thin pelecypod shells	Dk. yellowish brown
2B	Fine sand; a few shell fragments	Mod. yellowish brown
2.5A	Silty clay; abundant shell fragments	Dk. yellowish brown
2.5C	Very fine to coarse sand; shell fragments	Mod. yellowish brown
3A	Fine to medium sand	Grayish orange
3B	Silty clay; shell fragments	Mod. yellowish brown
3.5A	Silty sand	Mod. yellowish brown
3.5C	Silty sand	Dk. yellowish brown
4.5B	Silty clay; abundant shell fragments	Olive gray
5C	Silty sand; some shell fragments	Dk. yellowish brown
6B	Silty clay; a few shell fragments	Lt. olive gray
7B	Silty clay; a few shell fragments	Dk. yellowish brown
7.5A	Very fine to med. sand; shell fragments	Mod. yellowish brown
7.5B	Silty clay; some shell fragments	Lt. olive gray
7.5C	Very fine sand; some shell fragments	Olive gray
8A	Silty sand; many shell fragments	Olive gray
8.1A	Fine to coarse sand; shell fragments	Yellowish gray
8.5B	Silty clay	Lt. olive gray
9B	Silty clay; a few shell fragments	Lt. olive gray

* Color of dry samples based on National Research Council Color Chart.

400° C, and 14 Å structure having collapsed to a 10 Å structure.

Six samples contained amorphous or very poorly crystallized material in such large amounts that the crystalline clay minerals were masked by it. This material was removed by boiling each sample for 10 minutes in 1 N HCl and then in 1 N NH_4OH, each sample being washed thoroughly after each treatment. To determine the approximate amount of amorphous material present, an artificial diluent, composed of a 3:1 mixture of rapidly dried silica and alumina gels, was added in measured amounts to each cleaned sample until an X-ray pattern was obtained which matched the original cloudy pattern. The content of material amorphous to X-rays was surprisingly high, ranging from 50 to 75 percent by weight (table 4).

DIAGENETIC TRENDS

The fresh water portion of the Neuse River contains kaolinite, illite, montmorillonite, "chlorite," and mixed-layered aggregates (Brown and Ingram, 1954). Kaolinite is the dominant clay mineral. Illite occurs sporadically throughout the length of the stream. Montmorillonite occurs in small amounts in the upper part of the stream. Mixed-layered aggregates, possibly of illite and chlorite, occur throughout the length of the stream but increase in abundance downstream. A chloritic material similar to the "chlorite" in the estuary is present in the lower part of the stream. It

TABLE 4.—*Approximate amounts of amorphous material in samples giving "cloudy" patterns*

Sample No.	Amorphous Content
R4	75%
2.5C	75%
3.5C	65%
7.5A	60%
8A	50%
8.1A	55%

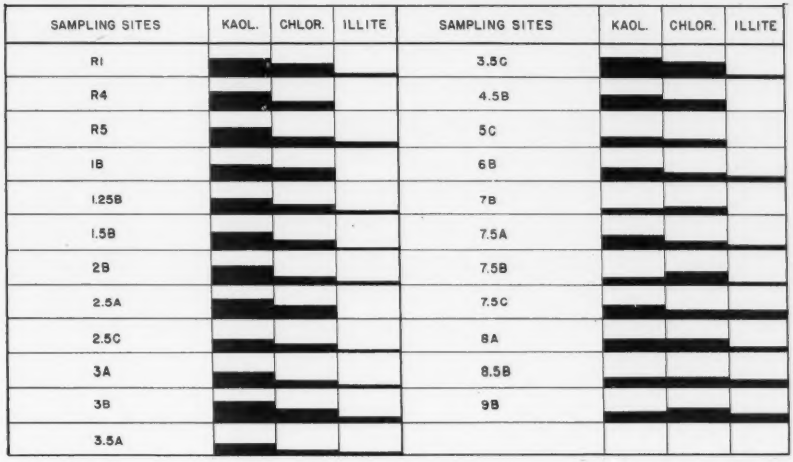

FIG. 3.—Intensities of basal 14 Å, 10 Å, and 7 Å lines from chlorite, illite, and kaolinite in Neuse River Estuary samples.

seems very probable that this "chlorite" was formed in an estuarine environment, either in the present estuary and washed upstream during times of high winds or in a past enlarged portion of the estuary.

In the downstream direction in the estuary, "chlorite" increases relative to kaolinite, and illite increases relative to both kaolinite and "chlorite." These trends are shown in figure 4 in which the ratios of the visually estimated intensities of the (00*l*) lines are plotted against the position in the estuary. In order to obtain smoother curves, running averages of five stations were used in the plot, with the average intensities being plotted at the average location of each group of stations. Three curves are plotted: "chlorite"/kaolinite, illite/"chlorite," and illite/kaolinite. The greatest amount of change in the curves of figure 4 lies between stations five and seven, just below the big bend in the estuary (see map, fig. 2). Apparently the bend has a constricting influence upon the inflow of sea water and diagenetic changes are retarded upstream from the bend.

It is realized that the use of intensities of (00*l*) lines as measures of absolute clay mineral abundances is open to many uncertainties; but as all the samples were handled in the same manner and as only ratios of intensities of (00*l*) lines in the same pattern were used, the results shown in figure 4 are considered to be significant.

Thus, kaolinite, the dominate clay mineral being introduced into the estuary, is being replaced gradually downstream by "chlorite" and illite. Other studies have shown analagous results. Grim, Dietz, and Bradley (1949) found illite to be the chief clay mineral forming in the Pacific Ocean off the California coast. Powers (1953) found "diagenetic chlorite" to be one of the major components of the clays of Chesapeake Bay.

The formation of "chlorite" and illite in the estuary can be explained in several different ways: (1) The two-layered kaolinite being introduced into the estuary could be changed into three-layered "chlorite" or illite by (a) the addition of a silica sheet to the kaolinite,

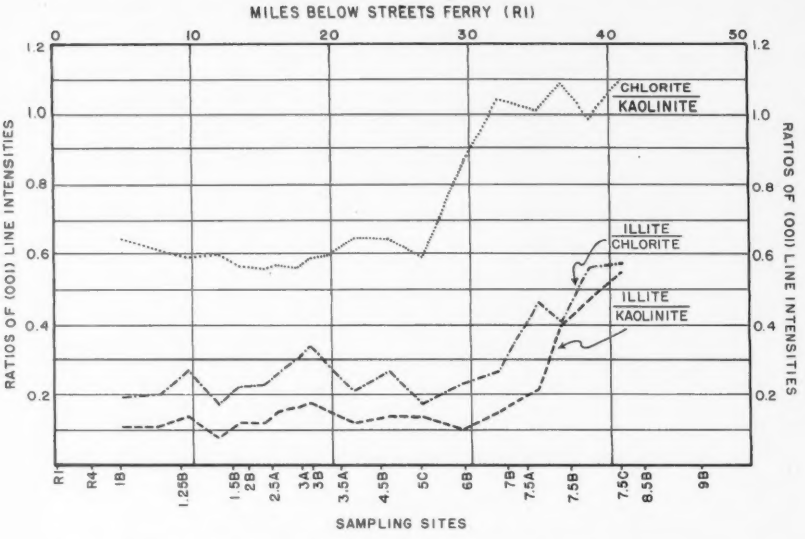

FIG. 4.—Ratios of intensities of basal (00*l*) clay mineral lines in Neuse River Estuary samples.

or (b) the breakdown of the kaolinite in the marine environment and the rearrangement of the material into "chlorite" or illite. (2) Much of the material amorphous to X-rays being washed into the estuary could be degraded soil and stream three-layered structures which are reconstructed in the marine environment. (3) Much of the material amorphous to X-rays could be truly amorphous materials brought in by the Neuse River and formed in the estuary from the clay minerals that are unstable under marine conditions. These amorphous materials could then enter into reactions that form "chlorite" and illite. The writers do not feel that the available evidence enables them to choose which one or ones of these mechanisms is most likely.

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U. S. COAST AND GEODETIC SURVEY, 1939, Neuse River and upper part of Bay River: Chart 538.

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FIGS. 1–2, SEPTEMBER, 1955

SANDSTONE CLASSIFICATION: RELATION TO COMPOSITION AND TEXTURE

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ABSTRACT

Composition and texture are two interdependant properties upon which any sound classification of sandstones must be based. The extent to which a sediment approaches maturity is measured by the complexity of these properties. Sandstone types are defined in terms of the positions they occupy relative to the crushed-rock complex and the perfectly mature sediment. These positions reflect the influence of both source terrane and the basin of deposition.

INTRODUCTION

Four recent papers dealing with the characteristics of sandstones and the problems relating to their classification (Dapples, Krumbein, and Sloss, 1953; Folk, 1954; Packham, 1954; Pettijohn, 1954) present somewhat divergent views and raise a number of provocative points. The present paper is based on a review of these articles and the older literature on the subject as well as mineralogical analyses of 61 sandstones, assembled largely from the literature.

BASIC PROPERTIES

"A rock . . . has only two basic, fundamental properties: composition and texture . . . " (Krynine, 1948). They reflect the source or provenance of the sediment, the extent to which it approaches maturity, and the fluidity of the depositing agent (Pettijohn, 1954). Any adequate scheme of sandstone classification must be based on both of these properties.

The composition of most sandstones may be fully stated in terms of the percentages of quartz, feldspar, metamorphic rock fragments, and clay. Figure 1 depicts the percentages of these constituents in 61 sandstones assembled by the writer.

Quartz and clay are the stable end-products of parent rock disintegration, whereas feldspar and rock fragments are metastable in both the source and depositional sites. Their presence in a sandstone indicates both rapid erosion and deposition. The source-rock index (Pettijohn, 1954) indicates the relative importance of feldspar-bearing and metamorphic terranes as contributors to the sediment. The maturity index indicates the extent to which either pre-depositional weathering, elimination of certain elements during transport, or derivation from a somewhat mature sediment have caused the sediment's composition to differ from the crushed-rock complex (Dapples, Krumbein, & Sloss, 1953). The relative importance of these three factors is generally difficult to assess.

Clay must be considered a textural rather than a compositional property, as it reveals nothing concerning the source rock but is an indicator of the depositional environment.

The texture of a sand may be considered on three levels: (1) presence or absence of clay; (2) sorting of the sand fraction; and (3) roundness of the grains. The presence or absence of clay is an index of the fluidity (density and viscosity) of the depositing agent, and the sorting and roundness of the sand fraction reflects the amount of reworking the sediment has been subjected to during one or more cycles of deposition.

Composition and texture are interdependant properties, a fact which is

particularly clear at certain levels of maturity. Consequently, they are not susceptible to separate treatment, as Folk (1954) suggests. To treat them as such leads to a classification which is meaningless in the sense that many of the "pigeonholes" are unoccupied in actuality and that texturally and genetically unrelated sandstones are placed in the same or similar categories.

BASIN OF DEPOSITION

Sandstones may be grouped into three main environmental categories, corresponding to three energy levels, on the basis of their textures: (1) the undisturbed environment (typical of rapid burial in the eugeosyncline); (2) the partially reworked zone (unstable shelves and intracratonic basins); and (3) the zone of intense reworking (stable shelves).

A considerable amount of interstitial clay characterizes sands deposited in the undisturbed zone. This indicates a depositional mechanism of high density and viscosity, presumably turbidity currents, which carried the sand out into this otherwise inaccessible zone. Essentially no energy was subsequently expended on the sediment to modify its initial structures. Quartz, feldspar, and metamorphic rock fragments may be present.

Sands deposited in the zone of moderate reworking are characterized by the absence of appreciable amounts of clay, but a poorly sorted sand fraction and feldspar and rock fragments (if available in the source area) are typical. Either the fluidity of the depositing agent was high (most probable) or, if it is low, enough reworking occurred to eliminate most of the clay which was present initially. This represents a second energy level, on which enough energy has been expended on the sediment to remove the clay but not to sort markedly the sand fraction or to destory or selectively remove unstable components.

The zone of intense reworking is one of high fluidity and energy input. Consequently, no clay is present, the sand fraction is well sorted, and unstable mineralogical components have been eliminated. Quartz (and chert) is the dominant mineral present. The products of this zone of sedimentation may be divided into two classes on the basis of the roundness of grains. Both are well sorted and high in quartz but the more mature sands, such as the St. Peter, are also highly rounded. It has been suggested (Dapples, Krumbein, & Sloss, 1953) that the less well rounded type is characteristic of the unstable shelf environment.

These environmental groups correspond in a general way to the "four stages of textural maturity, depending on the stability of the depositional site and the input of modifying energy" outlined by Folk (1954). Packham (1954), however, has questioned the validity of ascribing these stages to a single evolutionary sequence, and suggests that the bulk of the highly argillaceous sandstones belong to a separate, truncated series, due to their deposition in deep water and consequent inaccessability to reworking (traction) currents.

SOURCE

The areas supplying detritus to these various depositional sites may range from terranes which contribute a mechanically derived, crushed-rock complex to a deeply weathered laterite or an unmetamorphosed sedimentary terrane, both of which produce little but quartz and clay. This is largely a function of the tectonism of the region. The farther removed a sediment is from the crushed-rock complex the less reliable is its mineralogy as an indicator of provenance.

CLASSIFICATION

In order to test these concepts the mineralogies of 61 sandstones have been plotted on the accompanying triangle diagrams (figs. 1, 2). In figure 1 feldspar and metamorphic rock fragments, both metastable components, have been com-

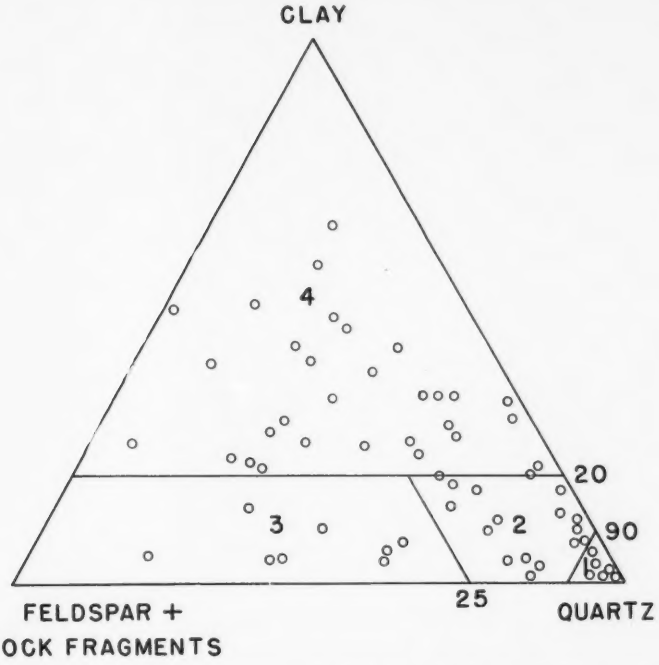

FIG. 1.—Plot of the four primary components of 61 sandstones. The numbers designate the following fields: 1—orthoquartzite; 2—protoquartzite; 3—arkose and lithic sandstone; 4—graywacke.

bined at one apex. Figure 2 also differs from most diagrams which have been used to date in that only three of the primary constituents are represented instead of combining some two in order to depict all four.

Despite Folk's recent (1954) attempt to redefine it, long usage (Pettijohn, 1943) has established the term "graywacke" as applicable to those sands high in interstitial clay, which are typically deposited in the undisturbed zone. The mineralogy of this type has been a matter of some debate (Folk, 1954; Krynine, 1948; Pettijohn, 1943, 1949; Tallman, 1949). The distribution of points on figure 2, marked by the prominent blank area between the two lateral concentrations, indicates that abundant feldspar and metamorphic rock fragments rarely occur in the same sandstone. This type of distribution also characterizes a plot of quartz, feldspar, and metamorphic rock fragments. Consequently, there must be two graywacke subtypes, the feldspathic and the "lithic" (Pettijohn, 1954). These correspond more or less closely to Krynine's (1948) high- and low-rank graywackes and Pettijohn's (1949) graywacke and subgraywacke. Although figure 2 does not indicate the actual percentages of feldspar and metamorphic rock fragments, they occur in appreciable quantities in most graywackes, and "quartzwackes" (Williams, Turner, and Gilbert,

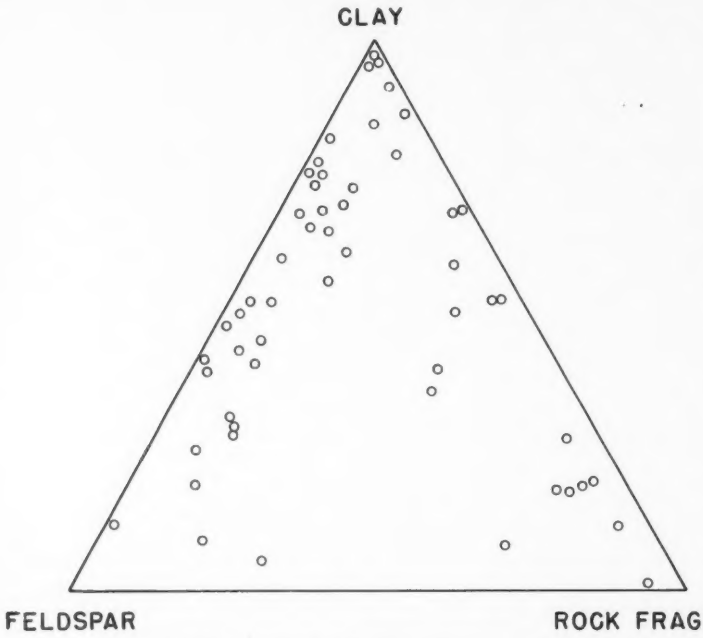

FIG. 2.—Plot of the three minor primary components of 54 sandstones, recalculated to 100 percent. Note the prominent blank area between the feldspar and metamorphic rock fragment poles. (Quartz comprises less than 90 percent of the original rock in each case.)

1954) are probably rather rare.

At the level of moderate reworking this compositional dichotomy still prevails but little or no clay is present. Consequently, rocks which may be of identical mineral composition as those in the undisturbed zone are no longer called graywackes but are termed arkoses and lithic sandstones. Although Dapples, Krumbein, and Sloss (1953) say that, " . . . no gradation is to be expected between arkoses and true graywackes. . . ," certain of the samples (Edwards, 1947; Briggs, personal communication) are indeed transitional, containing 20–40 percent each of feldspar and clay.

In the intensely reworked zone only quartz remains, and the rock clearly is an orthoquartzite. To call such a sand a graywacke if it contains "stretched metaquartzite" particles (Folk, 1954) is to ignore completely the textural evolution of the sediment.

On neither of the two diagrams are all four basic sandstone types depicted, figure 1 having a composite field and figure 2 lacking the orthoquartzite pole. It is an inescapable fact that no simple, two-dimensional diagram can show all four rock types. Most attempts to circumvent this fact have led to hybrid groupings which have obscured certain significant relationships. The prism of Dapples, Krumbein, and Sloss (1953) is an improvement but still has the defects that rock fragments and clay are combined at one apex and that it fails to account for the feldspathic graywackes.

A better solution might have been to place clay content on the vertical sorting axis (where it would have been restricted to the lower levels). Krynine (1948) and Folk (1954) are also in error in combining metamorphic rock fragments and clay, for, as Pettijohn (1954) points out, the former is a labile and the latter a stable component of the sediment. Also, the idea implicit in this grouping that the clay paste is a micaceous "hash" derived solely from the mechanical destruction of metamorphic rock fragments is questionable. Much of the sericite and chlorite which characterizes these pastes is undoubtedly the product of secondary organization of the original clay. Pettijohn's (1949) diagrams fail to account for the rock fragments present in many sandstones, and Folk (1954) errs in attempting to differentiate "metamorphic quartz" and place it with the rock fragments.

Figure 1 is an attempt to eliminate these undesirable features by combining feldspar and metamorphic rock fragments, both metastable components, at one apex. Nanz (Pettijohn, 1954) and Packham (1954) have also done this. The various fields in figure 1 correspond to the four primary (and one secondary) sandstone types. They are defined in the following manner: orthoquartzite—over 90 percent quartz (and chert); graywacke—over 20 percent clay; arkose and lithic sandstone—over 25 percent labile particles, more than half of which are, respectively, feldspar or metamorphic rock fragments; protoquartzite—less than 90 percent quartz, 20 percent clay, and 25 percent labile components. The protoquartzite group may be subdivided into feldspathic, argillaceous, and lithic subtypes, depending on which minor constituent predominates. These replace the terms feldspathic sandstone, subgraywacke and protoquartzite (as used by Pettijohn, 1954).

The classification here discussed is essentially the same as that recently presented by Pettijohn (1954). Graywacke and orthoquartzite are considered to be the two polar members of the sandstone family, the former representing the crushed-rock complex and the latter the absolute value on which all sediments converge in their journey toward maturity. This is not to imply, however, that the process of sediment evolution can always approach the limiting value. As Dapples, Krumbein, and Sloss (1953) and Packham (1954) point out, certain sands by virtue of their composition or depositional environment may be terminal deposits. Hence, the writer's objection to the term "subgraywacke" which implies derivation from a graywacke. Arkose and lithic sandstone are parallel (on the same level of maturity) intermediate types which are related much more closely than their diverse mineralogies would indicate. The failure of previous workers to recognize this similarity arose from the untenable assumption that the two rock types reflected drastically different tectonic states of the source area. High tectonism in the source (so the argument ran) would inevitably produce a flood of arkosic material, whereas the more mildly activated sources bordering geosynclines and intracratonic basins would yield only metamorphic rock fragments. Numerous examples can be cited to show, however, that a high degree of tectonism may never lead to the unroofing of the crystalline basement, and that lithic sandstones may be postorogenic deposits in the same manner as wedge arkoses. Likewise, blanket arkoses have their counterparts in blanket lithic sandstones. Protoquartzite is a secondary type which includes all sandstones transitional between arkose and lithic sandstone and orthoquartzite. As such, it is probably one of the most abundant sandstone types.

The classification is genetic in the sense that it is based on textural and compositional properties which reflect the position of each type with respect to the end members of the maturity scale. These hypothetical limits are the

crushed-rock complex, a sand derived from an undifferentiated source upon which a minimum of modifying energy has been expended, and the perfectly mature sediment, the result of complete lithologic convergence, accomplished by the expenditure of an infinite amount of energy. Perhaps a genetic classification can never be formulated in terms less abstract than these without descending to the level of mere detailed environmental ennumeration.

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FIGS. 1–4, SEPTEMBER, 1955

THE NATURE OF CORROSION ZONES IN THE MIDDLE ORDOVICIAN OF MINNESOTA[1]

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ABSTRACT

So-called corrosion zones are very common in the sedimentary rocks of the Upper Middle Ordovician in southeastern Minnesota and adjoining areas. They are usually represented by more or less irregular corrosion surfaces with a microconglomeratic covering. Both surfaces and covering have been produced by the mechanical erosion and chemical solution of the sediment on the sea floor through the action of bottom currents.

INTRODUCTION

So called corrosion zones are typical features of the upper part of the Middle Ordovician in southeastern Minnesota (Stauffer and Thiel, 1941). The nature of these structures and therefore their stratigraphic value are, however, still not fully understood. The following paper attempts to present data which may help clarify this problem.

The discovery of the corrosion zones seems to have been made by Chamberlin, (1882) in Wisconsin and Sardeson (1897) in Minnesota. In the following publications Sardeson (1898, 1906, 1916, 1926, 1937) described about eight such zones, one of which he described in detail from a genetical standpoint (1914). Further, the origin of the corrosion zones was discussed by Stauffer (1925), Pettijohn (1925, 1926, 1949), Thiel (1935, 1937) and Krumbein (1942). Some information on the corrosion zones was collected also by Elder (1936), Schwartz (1936), Stauffer and Theil (1933, 1941), Thomas (1937), Weiss (1953, 1954), and Majewske (1953).

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CHARACTERISTICS OF CORROSION ZONES

The term "corrosion zone" is used here to indicate certain parts of sediments showing traces of corrosion—corrosion surfaces, microconglomerates, and perhaps some layers of "brassy oolites." Such zones are located at different levels in the normal marine sediments of the Upper Middle Ordovician in southeastern Minnesota (fig. 1).

Corrosion surfaces are more or less parallel to the bedding and are either slightly uneven, somewhat pitted flakes (fig. 3b) or quite irregular, strongly pitted surfaces (fig. 2). They are covered and penetrated with different fucoidal trails and burrows at many places and have incrustations of bryozoa, crinoids, etc. at other places (Sardeson 1898). Numerous small pits five to six mm deep, probably made by worms, are especially typical for the corrosion zone at the top of the "Buff" submember in the Twin Cities area (fig. 3). Most of the corrosion surfaces have a black stain of finely divided iron sulphide up to more than one cm thick, which grades downward into the normal type of rock. Yellow pyrite crusts and segregations may occur together with pyritic strains. There are, however, some corrosion surfaces more or less free of black stain. In outcrops of the Glenwood formation corrosion zones show a limonitic incrustation, but at some places pyrite is present in

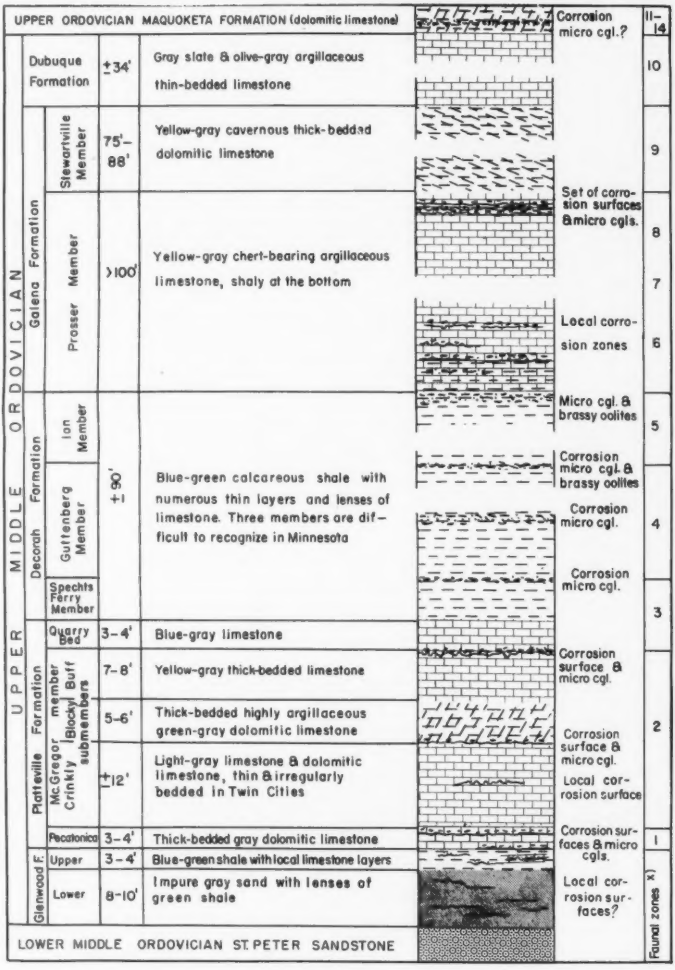

FIG. 1.—Geological column of the Upper Middle Ordovician in Minnesota, showing location of corrosion zones. Lower part of column according to outcrops in the Twin Cities. Upper part is projected from southeastern Minnesota and is absent in the area of the Twin Cities.

less weathered parts of the sediments (Majewske, 1953). Limonite therefore seems to be a product of the secondary oxidation of pyrite.

The corrosion surfaces are usually overlain by microconglomeratic beds[2] up to 10 cm or more thick composed of a calcareous, dolomitic, silicified, or shaly matrix containing coarse grains of quartz, phosphorite grains, and larger pebbles and slabs of limestone. The upper boundaries of microconglomeratic beds are more or less gradational. The limestone pebbles (fig. 3a) up to seven cm or more long and three cm or more thick lie parallel or inclined to the bedding whereas the larger limestone slabs tend to be parallel to the bedding. These slabs are up to 40 cm in length but only two-three cm thick at some places. The surface of pebbles and the upper surface of slabs may be black-stained and penetrated by fucoids. The phosphorite grains are especially characteristic of some corrosion zones and were described in detail by Pettijohn (1925, 1926). They are mostly small (± three mm) and irregular with black shiny surfaces and dull gray interiors. In the literature such grains have been called "pebbles," but they are of non-clastic origin and are smaller than pebble size. Neither is the term "nodule" applicable because many of the grains are organic fragments or molds.

According to Sardeson (1898) the corrosion zones have a tendency to occur close to the upper boundaries of the faunal zones (Sardeson 1897). They may occur, however, within such units also; for example, a set of corrosion zones occur within the Pecatonica member of the Platteville formation in Fillmore County (Weiss, 1953, 1954).

The oldest corrosion zones, in the form

[2] Because of the relatively small number of true pebbles in the rock previously called "corrosion conglomerate," the term "microconglomerate" is used here to point out the peculiarity of the sediment.

FIG. 2.—Strongly uneven corrosion surface covered with pyrite stain. Pyrite occurs as dark pigment and yellow segregation within the stain. Drill core of Platteville (?) limestone at Mason City, Iowa. Photographed by Photographic Laboratory of University of Minnesota.

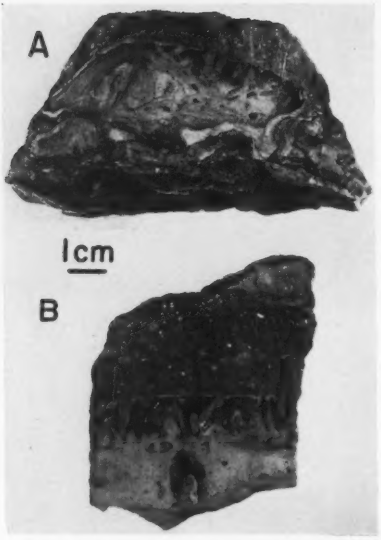

FIG. 3.—Corrosion zone near the boundary between Quarry bed and Buff limestone, Minneapolis, West River Road near St. Mary Hospital. a) Limestone pebble with dark stain and "worm burrows" from microconglomerate. b) Dark stained corrosion surface penetrated by numerous "worm holes." Some holes penetrate the stained zone. (Photographed by Photographic Laboratory of University of Minnesota.)

of limonite seams, occur in the Glenwood formation (Thiel 1935, 1937). Four or five such zones were observed by the writer in a sandstone quarry north of Minneapolis in Columbia Heights. At one place two of the limonite seams are strongly folded and broken although the strata above and below are undisturbed (fig. 4). The folding may be a result of a mud slide on the former sea floor. A few cracks in the corrosion zone at the top of the Buff submember at St. Mary hospital in Minneapolis may be of the same origin.

Corrosion zones in the Decorah shale are somewhat different. In this formation microconglomerates may occur without continuous corrosion surfaces at the base, and layers of "brassy oolites" are commonly present. These are 1½ to two mm in diameter and show well developed concentric structure and after partial solution in HCl leave opaline skeletons. Their x-ray pattern indicates goethite.

Corrosion zones such as are discussed here are widely distributed also in neighboring areas. They have been reported in Wisconsion, Iowa, Illinois, and Tennessee.[3] The writer observed several well developed corrosion zones in a drill core from Mason City, Iowa (fig. 2).

The horizontal extent of a single zone seems, however, to be variable. Some of the zones continue for miles, while others are very local. Such a local zone was reported, for example, by Pettijohn (1925) in a limestone lens in the Glenwood formation, while another one was found by the writer in the Crinkly submember in the Johnson Street Quarry in Minneapolis. The latter zone was about 20 feet long. The number of zones within a given formation seems to increase toward the south and southeast of Minneapolis. In Fillmore County in south-

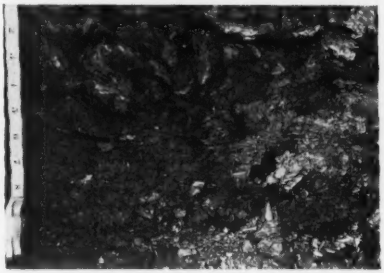

FIG. 4.—Intraformational folding of two Limonite seams and shales within Glenwood formation. Nick Helm sand quarry, Columbia Heights, North Minneapolis. From Majewske (1953).

[3] Agnew, 1948; Agnew and Heyl, 1946; Bain 1905, 1906; Bassler 1932, Plate 32A, B; Bays 1938; Bays and Raash 1935; Chamberlin 1882; Grant and Burchard 1907; Kay 1929; Kay 1935; Kay and Atwater 1935; Pettijohn 1925, 1926; Sardeson 1898; Templeton and Willman 1952, Fig. 15a, 16; Trowbridge and Shaw 1916.

eastern Minnesota, Weiss (1954) reported corrosion zones in groups at several horizons.

ORIGIN OF CORROSION ZONES

The following processes may be responsible for the development of corrosion zones:

A. Solution in hard rocks
B. Corrosion in unconsolidated sediments:
   a. Subaerial, during periods of drying,
   b. Subaqueous because of:
      1. Non-sedimentation at the profile of deposition, attended by wave action, influence of sea weed, etc.
      2. Solution of sediment because of increase in H_2S content of sea water
      3. Corrosion of sediment by bottom currents

Morphologically some of the corrosion surfaces are similar to stylolitic "undulation seams." According to Stockdale (1926) such structures have been produced by solution in solid rocks. In the case of the corrosion zones discussed, however, this interpretation cannot explain: (a) the origin of the microconglomerate beds with limestone pebbles and slabs; (b) the penetration of corrosion surfaces by fucoids and worm tubes; (c) the penetration of residual pyritic stain by some cavities (fig. 3); (d) the development of irregular pockets and knobs in solid rock.

Furthermore, the idea of corrosion during short periods of drying cannot be accepted because the marine sediments above and below corrosion surfaces do not show any evidence of drying. The corrosion surfaces are not associated with dessication fissures, rain imprints, and similar subaerial structures, but are penetrated by fucoids and worm tubes. Furthermore, the occurrence of pyritic stain instead of limonite indicates a subaquaceous condition of origin. Any transgression strong enough to produce microconglomerates would destroy the fine corrosion relief. (Sardeson 1914).

According to Thiel (1935) (Cf. Bain 1906) the limonitic corrosion seams at the base of the Glenwood formation were developed "when the sediments were built up to the profile of deposition in a shallow sea where an oxidizing environment prevailed." However, since most of the corrosion zones in the overlying strata show pyritic accumulations, there is no reason to suppose a different origin for the zones in the Glenwood formation. It seems more likely, therefore, that the limonite seams in the Glenwood were developed by the secondary oxidation of pyrite referred to above.

It is not clear what agents may have produced the structures described at the profile of deposition. For example,wave and tide actions would produce ripple marks rather than pitted surfaces, and all of the destruction of sediments in this area appears to be mechanical, whereas the corrosion surfaces show the influence of solution.

According to Stauffer (1925) and Pettijohn (1925, 1926, 1949) the corrosion took place in an anaerobic environment (similar to the Black Sea) through the solvent action of H_2S-bearing waters on calcareous mud. H_2S solutions, however, are not good solvents of $CaCO_3$. Moreover, some of the corrosion zones are not attended by pyrite stain.[4] The corrosion surfaces may have been penetrated by worm tubes and other animal trails and incrusted by bryozoa, crinoidea, etc., which contradict the presence of strongly anaerobic condition during corrosion. Moreover, an increase in H_2S would seem to be more probable in quiet basins, but the type of corrosion zones seem to indicate an increase in water agitation.

[4] Almost all of the carbonate rocks in the section discussed contain only small amounts of pyrite. Probably the pyrite that is present was deposited within but not on the surface of the mud as a result of the decomposition of organic remains within poorly aerated fine-grained sediment.

There still remains an alternative explanation of the structures observed. It is the corrosion by bottom currents[5] which has long been known in oceanic basins as well as in shallow seas. Such currents are of various origins and may produce both mechanical erosion and chemical solution of bottom sediments.[6]

The processes involved in such current erosion in clastic sediments may be described as follows:

1) Sedimentation of fine mud with small admixtures of sand takes place while no currents are present and normal benthonic life thrives on the sea floor.

2) Weak currents produce some outwash of the finest and lightest grains during the sedimentation and thus produce slight reduction in thickness and increase of grain size of sediments.

3) Continued increase in velocity of currents causes an increase in the amount of outwash of fine sediment. Some coarse grains may have been brought from other areas by the stronger currents, and the environment becomes unfavorable for bethonic life.

4) A further increase of current activity causes by-passing and even erosion of sediments previously deposited. The coarse residual material, including pebbles, accumulates on the eroded surfaces. At this stage the conditions for benthonic life are most severe and a specific biocoenose of burrowing and drilling organisms develops to produce some protection against the movement of water.

A decrease in the velocity of currents produces similar changes but in the reverse order. Corrosion surfaces with some residium become covered with coarse sediments grading upward into fine mud. Undoubtedly the position and strength of the currents were changed at times to produce different corrosion zones or they may have changed horizontally to produce changes in one and the same corrosion zone. Some of the currents were spread over larger areas to produce extensive corrosion zones, while others were very local. On the other hand, if currents were not sufficiently strong to develop true corrosion surfaces, their activity was limited to the formation of microconglomeratic beds, while very strong currents might erode certain parts of older sediments. This may be the reason for the absence of some lithic units and the thinning of other units at different places. Such relationships have been described in the literature for both old and recent sediments (Cf. Andrèe 1908, Heim 1924, Lutyj-Lutenko 1951, etc.).

The solution of calcareous mud may have been taking place simultaneously with the mechanical erosion. Such solvent action would be favored by an increase in CO_2 in bottom waters (Prokopovich 1952a), and may have caused the development of a pyritic residual stain on the corrosion surfaces. The environment was dynamic and solution may have been taking place also in the uppermost part of the accumulating sediment to produce black stained zones. Such stain may be absent, however, or only poorly developed if the sediment was originally low in pyrite. On the other hand, the origin of "brassy oolites" may have been connected with the accumulation of larger amounts of residual iron-bearing gels on the sea floor where the movements of water were favorable for the development of an oolitic structure.

According to Sardeson (1897, 1898, 1914) the corrosion surfaces occur in the uppermost parts of faunal zones. Faunal changes of such scale may have been caused by facies changes, for undoubtedly changes in current activity were at-

[5] The first very brief attempt to explain the corrosion zones by current action was that of Sardeson (1898). Referring to observation of Ruedemann (1897-98) in the state of New York, Sardeson briefly referred to the corrosion zones similar to those described by the writer as the result of submarine erosion and corrosion.

[6] Agassiz 1888, Alexander 1934, Andree 1908, 1920, Lebour 1874, Nansen 1904, Shepard 1948, Sverdrup et al. 1942, Twenhofel 1932, Verill 1878, Willis 1910 and many others.

tended by facies changes. The scarcity of fossils and the occurrence on corrosion surfaces of fucoidal burrows, worm tubes, incrustations of bryozoa etc. are good evidences of a dynamic environment produced by water currents which eventually led to the development of a specific biocoenose on the sea floor. No less significant is the association of corrosion zones with phosphatic grains and the occurrence of "oil rock" in the corresponding Ordovician sediments. (Agnew & Heil 1946; Bain 1905, 1906; Grant and Burchard, 1907; Majewske, 1953; etc.) Both may have been formed in the areas of upwelling water currents (Prokopovich, 1952).

CONCLUSIONS

Bottom currents may have covered different areas at different times so that the time-correlative value of the corrosion zones is slight. Even in the case of corrosion zones occurring in similar stratigraphical positions in different areas there is still a possibility that the corrosion was produced not by the same current everywhere, but by several currents and at somewhat different times. The paleogeographic significance of these structures as an indicator of ancient current activity is, however, very great. Moreover, they support the idea of the presence of many more or less local breaks in the marine geologic column (Twenhofel 1932).

The phenomena discussed is closely related to so-called turbidity currents. There is no doubt that currents strong enough to erode the sea floor will contain a load of sediment produced by the erosion of sediments already deposited, i.e., they will become turbid waters. Transportation by currents accounts also for the occurrence of some sand grain in the microconglomeratic beds. The type of biocoenose indicates, however, a more or less "hard bottom" with very dynamic water conditions rather than rapid deposition of great loads from muddy waters. Moreover, the invasion of muddy waters would cause the sedimentation of microconglomerate beds rich in clastic material, but such is not the case in most of the corrosion zones. In sandy and shaly sediments such as occur in the Glenwood and Decorah formations, "turbidity currents" may have played a part in the formation of corrosion zones, but in limestone strata, chemical corrosion is an extremely important factor. The turbidity of currents in all cases is, however, probably only a secondary result of submarine erosion, rather than the cause of development of the corrosion zones. Though somewhat similar structures have been discussed in the literature as a result of turbidity currents (Kuenen 1951, Natland & Kuenen 1951, and others), the zones observed by the writer are more reasonably explained by the activity of various types of "bottom currents" rather than by typical "turbidity currents."

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FIGS. 1–4, SEPTEMBER, 1955

DIAGNOSTIC HEAVY MINERALS IN THE REGOLITH OF MASSACHUSETTS

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ABSTRACT

Heavy mineral analyses were made on samples of soils and unconsolidated sediments collected from various parts of Massachusetts. It was noted that the outcropping and underlying bedrock has exerted a strong influence upon the character of the heavy mineral suite, particularly in the highland regions of the state. The soils and unconsolidated materials of the Connecticut Valley showed similar heavy mineral assemblages. No significant variations in heavy mineral content with change in horizon were observed in the soil profiles examined.

Data recorded for certain samples from the northwestern part of the state show the influence of wind-blown soils material.

INTRODUCTION

A survey of the heavy minerals in the soils and soil materials of Massachusetts was undertaken in an effort to determine if such studies would contribute to an understanding of the origin and morphology of the soils in the state.

It was found that the occurrence of diagnostic heavy minerals in addition to textural studies (Colby, et al, 1952) gave enough pertinent background information about soil origins so that a re-evaluation of soil mapping in Massachusetts seems advisable. An improved classification would not only utilize total mineral composition but also textural, structural, and general geological considerations in grouping the soils of the state.

Massachusetts can be divided into five general physiographic regions from west to east, (fig. 1). The Housatonic-Hoosic River valleys, the Western Highland, the Connecticut Valley, the Eastern Uplands and the Eastern Lowlands. For the purposes of this paper the first two regions are grouped together and are referred to as Western Massachusetts. The dotted areas on the map indicate the locations of samples.

PROCEDURE

Approximately 100 locations and several hundred samplings were made throughout the state. Both auger and channeling techniques were used in the sampling. Most of these samples were either separated by sieving or by the pipette method. All of the fractions of twenty-five samples were subjected to heavy mineral analysis using bromoform as the heavy liquid medium. Slides were mounted in arochlor ($n = 1.660$). The $-.125$ mm $+.062$ mm size or its approximate equivalent was found to be representative of the finer sand fractions, and the $-.250$ mm $+.125$ mm size or its approximate equivalent was

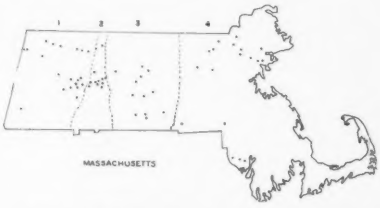


FIG. 1.—Geographic divisions used in this paper. 1. The Housatonic-Hoosic River Valley and the Western Highlands; 2. The Connecticut Valley; 3. The Eastern Uplands; 4. The Eastern Lowlands.

TABLE 1.—*Index to presentation of results*

Frequency	Approximate Percentage	Proportion
8+	90–100	
8	75–89	Very Abundant
8−	60–74	
7+	45–59	
7	35–44	Abundant
7−	28–34	
6+	23–27	
6	18–22	Fairly Abundant
6−	13–17	
5	7–12	Very Common
4	4–6	Common
3	2+–3	Fairly Common
2	1+–2	Scarce
1	½–1	Rare
x	< ½	Very Rare
–	—	Absent

From Evans, Hayman and Majeed.

found to be representative of the coarser sand fractions. In the case of tills which are bimodal the fractions in the region of the secondary maxima were used.

Throughout this paper frequency numbers (table 1) are used to denote the abundance of minerals, a scheme which is a slight modification of the one used by Evans, Hayman and Majeed (1933). These frequency numbers bear a logarithmic relation to the actual frequency as determined by counting, but the lower numbers depart somewhat from a true logarithmic plan. The advantages of this method of presenting data are that more emphasis is placed on the lower frequencies where a small change is more significant, and it obscures lesser fluctuations in frequency which fall within the range of experimental error in the techniques employed.

DISCUSSION

Table 2 gives the high, low, and average frequency numbers of the minerals in

TABLE 2

Mineral	Frequency											
	Western Massachusetts			Connecticut Valley			Eastern Uplands			Eastern Lowlands		
	H	L	A	H	L	A	H	L	A	H	L	A
Garnet	8−	–	6	7−	2	5	7+	3	6	7	5	6
Amphibole group	7−	–	6−	7	5	6+	7+	2	6−	7+	5	7−
Epidote group	8−	1	6−	7+	6−	7−	7	x	5	7	5	6
Black opaque	7+	3	6−	6	1	5	7−	1	5	7−	4	6−
Chlorite	6	–	3	7	1	5	4	–	2	6	–	3
Muscovite	7+	x	4	7−	1	4	4	–	1	6	–	3
Biotite	4	–	2	4	–	1	6	–	3	5	x	2
Sillimanite	1	–	–	2	–	1	8+	–	7−	5	x	3
Kyanite	2	–	x	2	–	x	5	–	1	1	–	0
Staurolite	6	–	3	3	–	1	1	–	x	2	–	x
Chlorotoid	6+	–	3	4	–	2	x	–	x	1	–	–
Zircon	6	–	3	4	–	2	6	–	4	5	x	3
Tourmaline	7	–	4	6−	–	2	4	–	1	4	1	2
Rutile	4	–	1	3	–	1	4	–	2	4	–	1

H = high; L = low; A = average for the samples examined from each region. Amphibole group includes hornblendes, actinolite, and tremolite. Epidote group includes epidote, zoisite, and clinozoisite. Black Opaque includes magnetite and opaque titanium bearing minerals.

the samples studied. These data show the quantitative distribution of the heavy mineral suite as it occurs in the Connecticut Valley, Western Massachusetts, Eastern Uplands, and Eastern Lowlands.

The average frequency of samples of soils from the Connecticut Valley appears as the dotted line in figure 2, and the associated nonsoil material as the solid line. The results shows that a close correlation exists between the soils and associated non-soils in this region with few exceptions. Epidote and tourmaline frequency averages are slightly higher for the soil samples. Garnet, amphibole, black opaque, biotite, chlorotoid, and zircon show slightly higher averages in the non-soils. More intensive studies are necessary to establish the extent of such correlations.

Profile charts of heavy mineral frequencies in relation to the locations of samples throughout the four regions of the state appear in figure 3. The frequency units of the −.125 mm +.062 mm fractions were plotted vertically and the samples by number were plotted horizontally. This gives a general overall view of the distribution of each mineral throughout the sampled areas of the state.

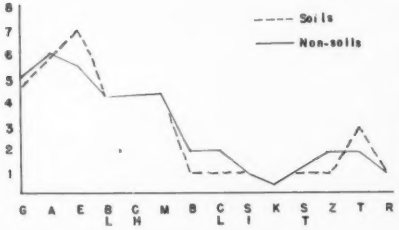


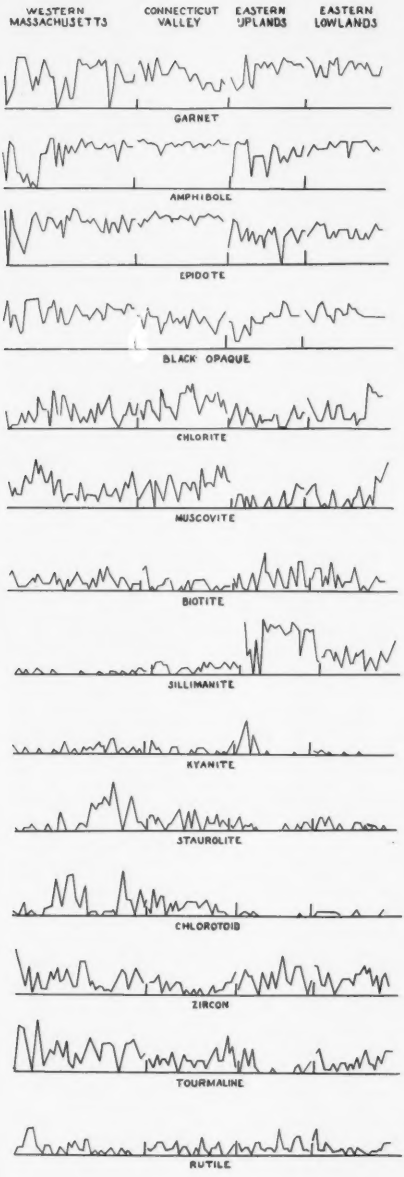
FIG. 2.—The average frequency of heavy minerals in the Connecticut Valley samples. The vertical represents the average mineral frequencies. The horizontal represents the minerals. G = garnet, A = amphiboles, E = epidote, BL = black opaque, Ch = chlorite, M = muscovite, B = biotite, Cl = chlorotoid, Si = sillimanite, K = kyanite, St = staurolite, Z = zircon, T = tourmaline, R = rutile.

In figure 3 garnet appears to be most abundant in eastern Massachusetts and shows considerable deviations in abundance in the western part of the state. Amphibole and epidote have a similar pattern and appear to be generally abundant throughout the state. The lowest frequencies are recorded in the eastern and western uplands. In the Connecticut Valley the high frequencies of both amphibole and epidote show fewest deviations. The highest concentration of black opaque is in the western uplands. Here again the upland samples show the largest variations in frequencies.

The most noteworthy feature of the chlorite and muscovite profiles (fig. 3) is the increasing concentration of these minerals in the Connecticut Valley and westward. Both minerals show wide deviations in frequency throughout the state. Biotite is generally less abundant than chlorite and muscovite, and its higher concentrations are found in eastern Massachusetts, particularly in the uplands.

The next four profiles below that for biotite are representative of the typically diagnostic heavy minerals. Other minerals, which may occasionally be diagnostic but generally were found in minor quantities in the samples collected are pyroxenes, hematite, andalusite, corundum, spinel, and monazite. The samples of chlorotoid in the profile indicate that it is scarce to rare in the eastern Massachusetts samples. It increases in quantity in the Connecticut Valley and becomes fairly abundant in some samples of western Massachusetts. In the western part of the state sillimanite is either absent or rare in the areas sampled. Its quantity increases slightly in the Connecticut Valley. In the eastern uplands it is present in abundance and it is common in the eastern lowlands.

Kyanite is typical of many minor minerals found in these samples. It is absent or scarce in most samples, with a few samples having higher concentra-



tions. Staurolite is fairly abundant in the central part of western Massachusetts, but at other locations sampled, it occurs with considerably lower frequency. It should be remembered that these profiles are of the −.125 mm size fraction. Kyanite and staurolite concentrations increase in the coarser fractions, particularly in the samples from western Massachusetts.

The last three profiles appearing in figure 3 represent the relative frequencies of the so-called stable minerals, zircon, tourmaline, and rutile. All three have their most prominent development in the upland regions. The zircon content is higher than the tourmaline in the eastern part of the state; the reverse is true in the Western Uplands. Rutile content is highest in the Eastern Uplands. The fact that common to abundant frequencies are recorded for all three of the stable minerals, particularly in the Western Uplands, does not indicate that the Western Upland soils have been subjected to a high degree of weathering. It is more likely to indicate that unusual amounts of these minerals have been derived from the outcropping rocks in the region.

This investigation has uncovered considerable evidence showing the strong influence of the bedrock on the sand fractions of the surficial deposits near outcrops, particularly in the highlands. The numbers on the map in figure 4 represent areas where significant frequencies of one or more minerals appear to be related to bedrock or to nearby outcrops.

Samples taken from the plots at the Massachusetts Agricultural Station in the Greylock area (fig. 4, loc. 1) show unusually high frequencies of tourmaline and black opaque with garnet scarce in the finer fractions. Underlying Greylock schist contains significant quantities of

FIG. 3.—Profile charts of heavy mineral frequencies (vertical) in the regolith of Massachusetts in their geographic setting (horizontal).

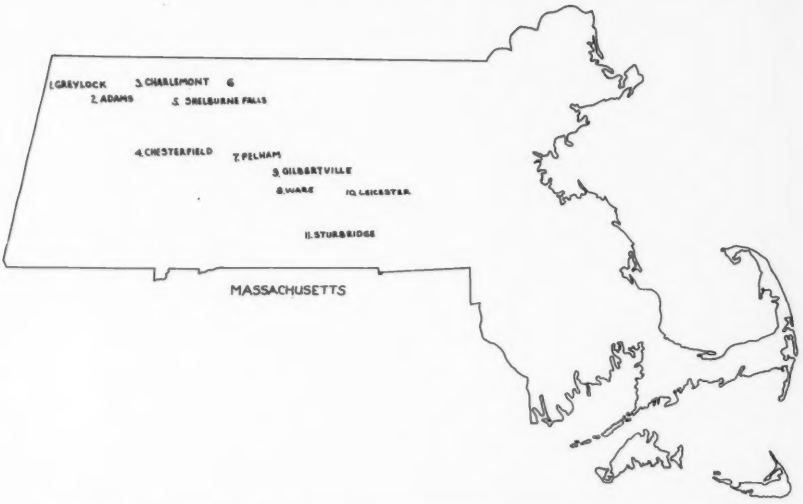


FIG. 4.—Areas where specific mineralogic correlations have been made with the bed rock.

tourmaline, ilmenite, and magnetite. No garnet is reported. Some of the magnetite in the soil samples shows octahedral habit even in grains which have been partially altered to hematite during the soil-forming process. Thus the evidence points to a residual derivation for this soil body. The garnet which is present in minor quantities may be representative of the material added to this soil by wind.

Near the town of Adams in the northwestern part of the state (fig. 4, loc. 2) another area was found having an extremely high concentration of tourmaline. The eastern part of the Berkshire schist contains much tourmaline, and outcrops of it are widespread in this area.

Samples taken northeast of the town of Charlemont (fig. 4, loc. 3) have abundant amounts of muscovite. The Savoy schist is the main outcropping formation of the area and contains a high percentage of muscovite.

In the Central Western Massachusetts region, (fig. 4, loc. 4) staurolite is common to very common and in places fairly abundant. The staurolite distribution indicates that the Conway schist, which is the most widespread formation in this region, obviously added to regolith material. In the Conway formation Emerson (1917) pointed out that there was a general abundance of staurolites to the south which became increasingly rare toward the north. This same pattern occurs in the surficial material covering the Conway schist. Zoisite and clinozoisite are the common epidote minerals in the bedrock of this region and in the vicinity of Shelburne Falls (fig. 4, loc. 5) and this too is reflected in the soils and soil materials.

Kyanite is generally scarce in the sampling throughout Massachusetts but is more common in the coarser fraction, particularly from the Western Uplands. Kyanite is reported to be common in such Western Uplands formations as the Conway and Goshen schists.

The high percentage of tremolite in the samples taken in the Great Barrington area of Southwestern Massachusetts reflects a characteristic of the underlying and outcropping Stockbridge formation.

South of the town of Northfield in the

Connecticut Valley region, (fig. 5, loc. 6), soil samples which have certain characteristics of loess have an abundance of chlorite and muscovite in both the coarse and fine heavy fractions. Surficial material taken just below these samples adhere more closely to the lower average frequencies which are more commonly found in the Connecticut Valley region. The rocks which outcrop here such as the Amherst schist, Devonian Bernardston formation, and the Monson granodiorite contain some high concentrations of chlorite and muscovite. While some of these minerals in the soils appear to be derived from such sources, the morphological characteristics of the soils also indicate an unusual influx of windblown material perhaps from the northwest.

Moving to the Eastern Uplands, north of the town of Pelham (fig. 4, loc. 7) several residual soil samples had a very high frequency of tremolite-actinolite forms of amphibole. Tremolite-actinolite (Emerson, 1917) are reported as common in the Northfieldite border phase of the Pelham granite which is the outcropping and underlying formation in this section. In the eastern uplands tremolite is not found in any large quantities where this formation is absent.

Garnet and sillimanite were found to be very abundant in samples taken in the highlands near the town of Ware (fig. 4, loc. 8). In the coarse fractions garnet is extremely frequent and in the finer fraction sillimanite is more frequent. This sampling area is underlain by the Coys Hill granite which is noted for its high concentrations of sillimanite and garnet.

Somewhat north of the town of Ware near the town of Gilbertville, (fig. 4, loc. 9) samples had an uncharacteristic scarcity of sillimanite. Significantly, the Hardwick granite which outcrops throughout this locality has no sillimanite. A similar situation is found in the eastern section of the uplands near the town of Leicester (fig. 4, loc. 10). Those samples where sillimanite was either absent or rare were taken above the so-called Paxton quartz schist which is non-sillimanitic in this area.

Biotite which is found in much of the Brimfield schist is reflected in the surficial formation in the vicinity of Sturbridge (fig. 4, loc. 11) where the Brimfield contains an abundance of biotite. The rutile content is highest in the Eastern Uplands. Outcrops of rutile bearing formations are reported in this area (Johansson, 1950).

The above examples serve as an illustration of the type of correlation possible between the mineral content of the surficial materials with the subsurface and outcropping bedrock. The findings indicate that the addition of material by the outcropping and underlying bedrock is much more extensive in highland samples than in those from lower areas. A corollary to this observation is that glacial action appears to be able to reduce coarse material to sand size fractions with relatively little transportation, along with the releasing of sand size particles. More intensive heavy mineral analyses could make a considerable body of such information available which would prove useful in explaining the origins of the soils as well as the variations that occur among them. Specifically the survey has proved helpful in determining the residual or transported nature of certain soils. Rudimentary beginnings have been made in understanding the paleogeography associated with the formation of these mineral grains and the results show that it is a fruitful field of study.

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SEPTEMBER, 1955

DIFFERENTIAL SURFACE WEATHERING OF BIGHORN DOLOMITE

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ABSTRACT

The Bighorn dolomite was examined to determine the cause of its uneven weathering surface. Samples collected from two localities at the southeast end of the Wind River Range in Wyoming were found to contain quartz and opal in small patches which account for the very rough outcrop surface.

INTRODUCTION

This study was made to determine the characteristics of the Bighorn dolomite which cause the rough, knobby surface upon weathering. Particular attention was given to the variations in color on fresh and weathered surfaces, because it was thought that the colors might be a reflection of mineralogical differences which could influence the rate and degree of weathering.

Fourteen samples were collected at two localities on the northeastern slope of the Wind River Mountains in west-central Wyoming. The first locality was on the north side of Wyoming Highway No. 28 about 20 miles south of Lander and four miles north of the Atlantic City turnoff; seven samples were taken from a 60-foot vertical section. The second locality was on the north wall of Sinks Canyon, Middle Fork of the Popo Agie River, about nine miles southwest of Lander; six samples were taken from a 60-foot vertical section. Each sample from each locality was taken from a different ledge. The ledges in both localities are about eight to 12 feet thick. One sample was taken from a large weathered boulder lying below the outcrop on the south side of Sinks Canyon. Each sample was a large piece weighing approximately 15 pounds.

The laboratory investigation of the samples included etching, staining, petrographic study of thin sections and insoluble residues, and X-ray powder diffraction analyses of the rock and of the insoluble residues of some samples.

The topographic expression of the Bighorn dolomite, its surficial appearance caused by differential weathering, and its appearance on fresh surfaces have been described by Darton (1906) and Blackwelder (1913). It occurs as massive layers which form steep cliffs along the outcrop. The exposed and weathered surface is one of depressions, sharp to knob-like protuberances, and irregular reticulate ridges. The most common colors of the fresh rock lie within the yellow-red and yellow hues of light value and low to medium saturation (Munsell Rock Color Chart), whereas the colors of the darker portions of mottled rock lie within the yellow-red and yellow hues of medium to light value and low to medium saturation. The colors of the weathered surfaces generally lie more nearly within the middle of the yellow-red and yellow hues of medium value and low to medium saturation than to the colors of the non-weathered counterparts.

PREPARATION OF SAMPLES FOR TESTING

The samples were sawed into slabs and some of the sawed faces ground to a nearly plane surface for etching. Some corresponding parts of the slabs were

cut out for thin sectioning. Other of the slabs were smoothed in order to be used in staining tests.

Fragments were broken from the large samples for solubility tests and to obtain insoluble residues. Other fragments broken from both weathered surfaces and the more nearly fresh interiors of the large samples were crushed and ground to fine powder for X-ray analysis.

ETCH TESTS

A solution of eight percent hydrochloric acid was used for etching. A tight dam of modeling clay was placed around a spot about two cm in diameter so that the acid was about one cm deep on the rock. The acid was left in contact with the rock for 10 minutes after which it was poured off and the rock thoroughly washed with distilled water. This procedure was used on both light and dark colored spots, on the most porous and least porous portions, on coarse-grained and fine-grained portions, and on parts of the rock of intermediate characteristics. All reacted similarly although the degree and rapidity of etching were greater in the more porous portions of the rock. There was no differential etching that could be attributed to mixed presence of calcite and dolomite, but the presence of disseminated silica was indicated.

STAINING TESTS

The technique using $Cu(NO_3)_2$ stain, described by Rodgers (1940), was used in an attempt to differentiate calcite from dolomite to determine their relative abundance. In this technique the calcite is stained blue, whereas the dolomite remains unstained.

The results of the tests indicated the presence of more calcite than later X-ray analyses showed. This discrepancy was due possibly to fineness of grain in most parts of the rock which made it difficult to distinguish individual stained grains; or, more probably, to high porosity and permeability which permitted absorption of reagents and therefore a show of too much color. This phenomenon was especially prominent in the weathered margins of specimens, along fractures, and in areas of grainy texture due to partial solution and recrystallization.

INSOLUBLE RESIDUES

A portion of each sample was placed separately in a six molar solution of hydrochloric acid and permitted to stand under a hood at about 30° C until no further reaction between acid and solid was noticable. More acid was added, until no further reaction was observed. The liquid was then evaporated slowly at about 40° C, and the residues were washed with distilled water and dried. No attempt was made to determine the percentage of insoluble material, but the mineralogy of the insolubles was determined by the X-ray spectrometer and microscopic examination.

All the samples contained insoluble materials, but the type and amount were variable. Quartz and opal were the most common, illite was present in minor quantity in most of the residues, and pyrite mantled by iron oxide occurred as occasional grains.

Except for occasional grains, the quartz occurred both as compact irregular masses and as fragile lacy bodies which filled the spaces between carbonate grains. All these masses were marked by a great number of dolomolds which indicated that calcite had been more easily replaced than had dolomite. The occurrence, and therefore the general appearance, of the opal was similar to that of the quartz except that the opal was translucent whereas the quartz was exceptionally clear and glassy. A very small percentage of oolitic chalcedony occurred in two samples.

The samples from the most prominent ledges, those of the lower part of the upper half of the 60-foot section in both localities, contained considerably more quartz and opal than those from more

recessed layers. Also, the samples from the prominent ledges contained less illite than those from the more recessed layers. Limonite and pyrite were most common in samples with the higher illite content.

PETROGRAPHIC STUDY OF THIN SECTIONS

The gross petrographic characteristics of the Bighorn dolomite in the two localities were very similar. Minor variations were caused primarily by the proportions of fine-grained and coarse-grained carbonate and by the presence or absence of quartz and opal. Illite occured in very small amount and could not be detected easily.

The compositions of the carbonates were not determined petrographically. The staining, X-ray, and etching tests indicated that both calcite and dolomite were present. Apparently the size of grain was not an accurate reflection of the composition of the carbonate, as was borne out by the staining tests.

Most of the quartz and opal had much the same mode of occurrence and habit, namely, as fine, disseminated irregular masses which were characterized by extremely fine projections from the main mass between adjacent grains of carbonate. Quartz, which showed "comb" structure, partially filled vugs and short fractures. Fossil fragments of several kinds, all of which presumably had been carbonate, were composed of clear quartz. These latter masses are, in general, the largest individual masses of quartz in these samples.

Estimated percentages of quartz and opal combined ranged from less than one to 15. Those samples with the higher percentage of quartz and opal definitely contained some of each mode of occurrence and habit. Those samples which had the higher percentage of fine, disseminated quartz, but not necessarily the highest percentage of all quartz, came from the more resistant ledges near the middle of the stratigraphic sections.

The minor constituents in the thin sections were limonite, muscovite, and a very fine, cloudy material which was assumed to be a clay mineral. This assumption was substantiated by X-ray diffraction patterns of the fine-grained insoluble residues which showed illite to be present. The muscovite is shredded and irregular in outline and is thought to be detrital.

The spatial distribution and boundary relationships between the fine and coarse carbonate and the quartz and opal are very irregular. Stringers and veinlets of coarse-grained carbonate penetrate into and cut across patches of fine-grained carbonate. Quartz and opal cut across both carbonates and protrude between and into grains of carbonate.

X-RAY POWDER DIFFRACTION ANALYSES

A North American Phillips X-ray Spectrometer with a copper tube was used. Fragments that were analyzed consisted, individually, of the light-colored rock, the dark-colored rock, and parts of the knobs on the rough weathered surface.

X-ray powder-diffraction diagrams of each of these types of material from each sample showed dolomite to be the chief constituent. Calcite was a very minor constituent. The patterns of almost every sample showed weak to strong peaks for quartz. In some samples only material from the knobs from weathered surfaces contained enough quartz to cause definite quartz peaks in the patterns. In general the patterns of samples from the southern end of the Wind River Mountains indicated more quartz than did those of the samples from Sinks Canyon. There was no distinction between the amount of quartz in the light-colored and the dark-colored portions.

The presence of illite was not noticeably indicated on the X-ray diffraction patterns of any of the rock samples. However, the patterns for several of the in-

soluble residues showed strong peaks for illite. None of these samples was rich in quartz or opal, and all were from the less resistant ledges that were sampled.

SUMMARY AND CONCLUSIONS

The weathered surfaces of the Bighorn dolomite, in the two localities sampled, are rough and knobby due to the presence of quartz and opal which are irregularly and unequally distributed. The parts of the exposed rock which contain an appreciable amount of quartz and opal are left as irregular shaped knobs and ridges separated by valleys and depressions of similar irregularity.

The cause of over-all greater resistance of the layers of dolomite in the lower part of the upper half of the section is the presence of a relatively higher quartz and opal content.

The mottled coloring of the dolomite apparently is not due to dolomite-calcite distribution nor to quartz and opal distribution. The cause was not determined.

Most of the quartz and all the opal are authigenic. Detrital grains of quartz are extremely uncommon.

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PL. 1, FIG. 1, SEPTEMBER, 1955

EXTRAORDINARILY COARSE EOLIAN SAND OF THE ICA DESERT, PERU

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ABSTRACT

Eolian sandstreams along the Peruvian coast locally are composed of well-sorted granules of country rock in which the modal class exceeds three mm and the coarsest grains are six or seven mm in diameter. Generally this coarse sand is a first generation deposit and does not owe its sorting to earlier generations of sediments.

The Pacific coast of Peru with an average precipitation of less than half a centimeter per year for more than a thousand miles south of Punta Aguja is one of the most arid regions on earth. This region contains many remarkable examples of sand dunes (Bowman, 1916; Broggi, 1952; Simons and Erickson, 1953). Sea breezes which strike the coast from the south and southwest are remarkably uniform in direction and maximum velocity although there are marked diurnal fluctuations. Maximum velocities rarely exceed 50 or 60 kilometers per hour (*fide* Dr. Jorge A. Broggi). In the morning the wind begins to blow from the south, veering to the southwest in the afternoon, dying down by evening. In many places the offshore wind is quite strong, blowing a gale around midday. It is these persistent sea breezes, more or less parallel with the coast, which are responsible for the Humboldt current and upwelling cold water (Schweigger, 1949). A characteristic effect of the dominantly offshore winds are great trains of dunes of fine sand (termed chiflones by Broggi, 1952), and smaller sand rows, "sand streams" (Simons and Erickson, 1953), and coarse ripples (pl. 1, fig. 1). These are linear features that clearly mark dominant jet streams of the wind. As the wind jets approach the Andean front, they generally are deflected upward and clockwise in accordance with the Coriolis effect.

From June to August, 1953, we were privileged, under the auspices of Seaoil, Limited, to make an examination of the geology along the coastal area between Pisco and Nazca, between approximately 15° and 16° south latitude in southern Peru. In this area "sand streams" and great ripples of coarse sand are characteristic of the more exposed areas of elevated terraces and upland surfaces of the Coastal Cordillera. Throughout this area the coast is marked by a high sea cliff. Beaches are so narrow that they supply little or no sand to the adjoining uplands. There are no intermittent streams of water in the upland areas and none of the coarse sand under consideration here can be attributed to waves and running water. It is directly a product of weathering and deflation.

As described by Simons and Erickson (1953), the sand streams consist of alternating light-colored ridges and darker furrows of sand, aligned more or less parallel with the predominant wind direction. These bands are crossed by coarse ripples. The sand is exceptionally coarse at elevated localities near the shore where wind velocities are highest. Diameters of the most abundant grains exceed three millimeters at many localities. The coarsest material observed by us was on the Pampa Corre Viento near Bahía Media Luna along the automobile road to Playa Media Luna, 95 kilometers west-northwest of Nazca (fig.

PLATE 1

FIG. 1.—View southwest toward Punta Zárate, near Otuma, Peru. Wind ripples in coarse sand. Distance between crest of ripples, one to one and a half meters; amplitude, 15–30 cm.

FIG. 2.—Eolian grains of fossil oyster-shell fragments, Talbo Bay, north of Punta Parácas, Peninsula Parácas, Peru.

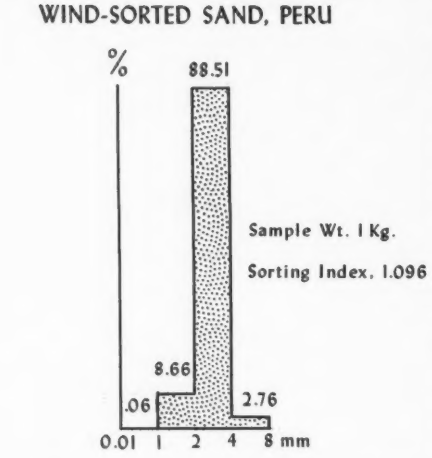

FIG. 1.—Histogram of very coarse sand from Pampa Corre Viento, Southern Peru.

1). In this place sand streams break down into irregular ripples of fine gravel in which the modal class lies between two and four millimeters in diameter. Excavation of the ripples (as elsewhere) shows that the granules are of uniform size to bedrock beneath the ripple. Fine sand lodges temporarily on the lee side of the ripples and between the coarse grains. The relief of individual ripples of coarse sand ranges from about .15 to .30 meters and the distance between crests is about 1.0 to 1.5 meters. Granules consist in most areas of rather well-rounded fragments of massive hypabyssal rocks of andesitic composition. Locally they are formed of fragments of fossil mollusks of Eocene age (pl. 1, fig. 2). In some cases the sand streams can be traced up inclined surfaces to the bedrock source. Elsewhere it is clear that the sand grains have been rolled for several miles from any possible place of origin. It seems clear that the grains are residual, wind-sorted, and wind-transported material from which the finer fractions have been selectively winnowed. The textures are not inherited from older sedimentary rocks. This material is transported along nearly horizontal surfaces and down slopes by rolling and saltation. It shows a high order of sorting characteristic of eolian sands.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 25, NO. 3, PP. 229–234
FIGS. 1–4, SEPTEMBER, 1955

SUMMARY OF "CONCERNING SOME ADDITIONAL AIDS IN STUDYING SEDIMENTARY FORMATIONS" BY M. S. SHVETSOV[1]

RICHARD D. TERRY AND GEORGE V. CHILINGAR
University of Southern California, Los Angeles, California

ABSTRACT

An article by the Russian sedimentologist, Shvetsov, is summarized for wider distribution among Western geologists. The summary includes reproduction of Shvetsov's diagrams for visual estimation of percentages of minerals in rock sections. In addition, the article points out the usefulness of water and acid for visual study of rock sections.

INTRODUCTION

Major factors which lead to lack of knowledge of Soviet research among Western sedimentologists are: (1) difficulty of locating Soviet publications, (2) inability to read Russian, and (3) the high cost of translations. In view of this situation, the authors believe that the article by M. S. Shvetsov (1954) is of enough interest to warrant an English summary and full reproduction of his diagrams.

SUMMARY OF SHVETSOV'S ARTICLE

Since percentages are useful in stratigraphic correlations and genetic studies, it is important to determine accurately the percentages of the components (sand grains, shells, etc.) in sedimentary rocks. Any significant error in estimating the various percentages will lead to erroneous interpretation and correlation. Numerous methods have been described for determining the amount of various constituents in rocks, but most geologists prefer visual estimation because it is less time consuming.

Shvetsov has made a comprehensive study on the accuracy of visual estimation methods. Part of the study consisted of asking each member of a geological meeting to estimate the percentages of various components in unmarked sections. He found that there was great variation in their results; some geologists greatly over- or underestimated, but there was a tendency toward overestimation—even reaching five to seven times the correct values. In addition, same observers gave different estimates for different orientations of the same rock. This was due to variations in the character and distribution of grains in rock sections of different orientation.

Shvetsov's contribution consists of constructing diagrams representing various percentages of grains (figs. 1–4). The diagrams can be used to compare with either hand specimens or microscopic rock slides. Shvetsov's diagrams were prepared by cutting out two circles of equal diameter—one white and one black. A sector equal to the desired percentage was then cut into small pieces and glued on the opposite colored circle.

It is generally known among geologists that wetting a rock section will usually bring out many details not otherwise seen in a dry section. However, it is necessary to maintain a water layer during the observation. Shvetsov points out the additional aid of first pouring dilute hydrochloric acid on a rock section followed by a washing with water. In addition to removing surface films and dust, the acid frequently etches the surface so that a low relief is formed due to different solubilities of the rock constituents.

[1] Allan Hancock Contribution No. 160.

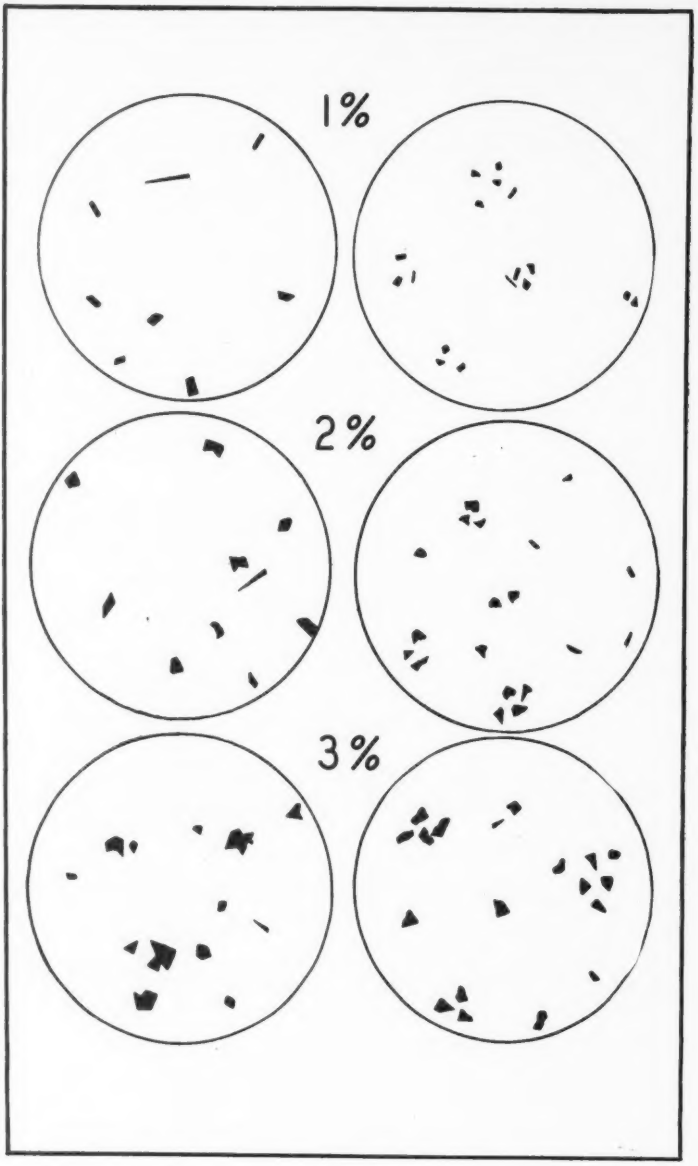

FIG. 1

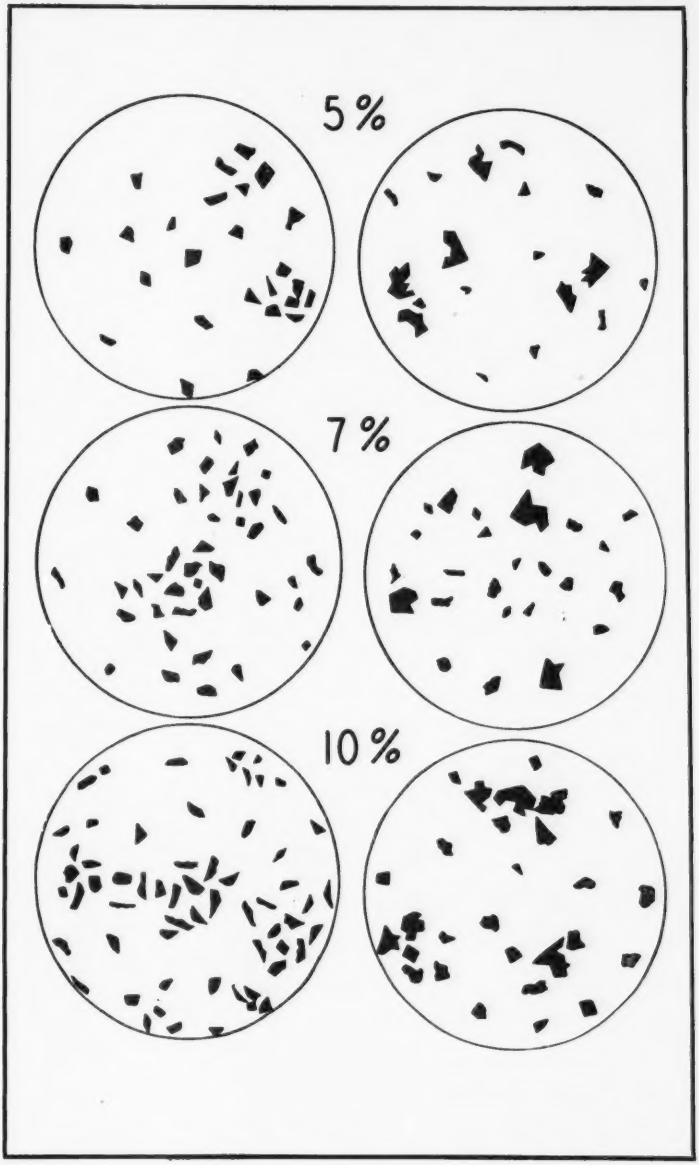

FIG 2

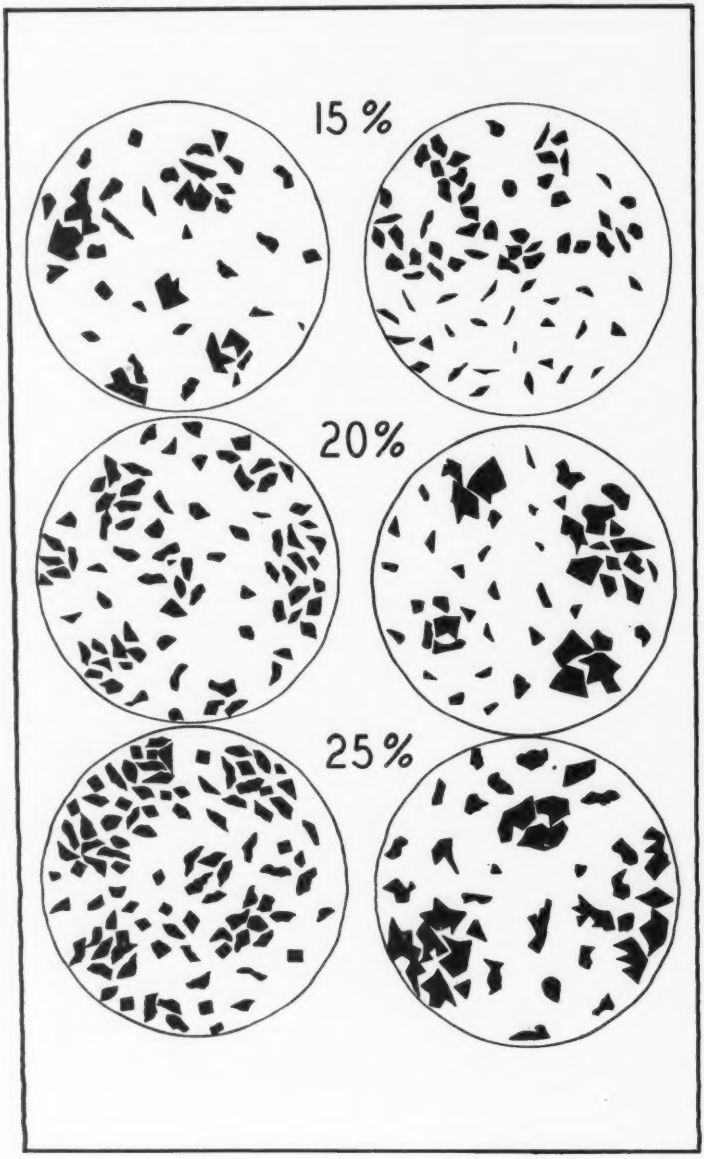

FIG. 3

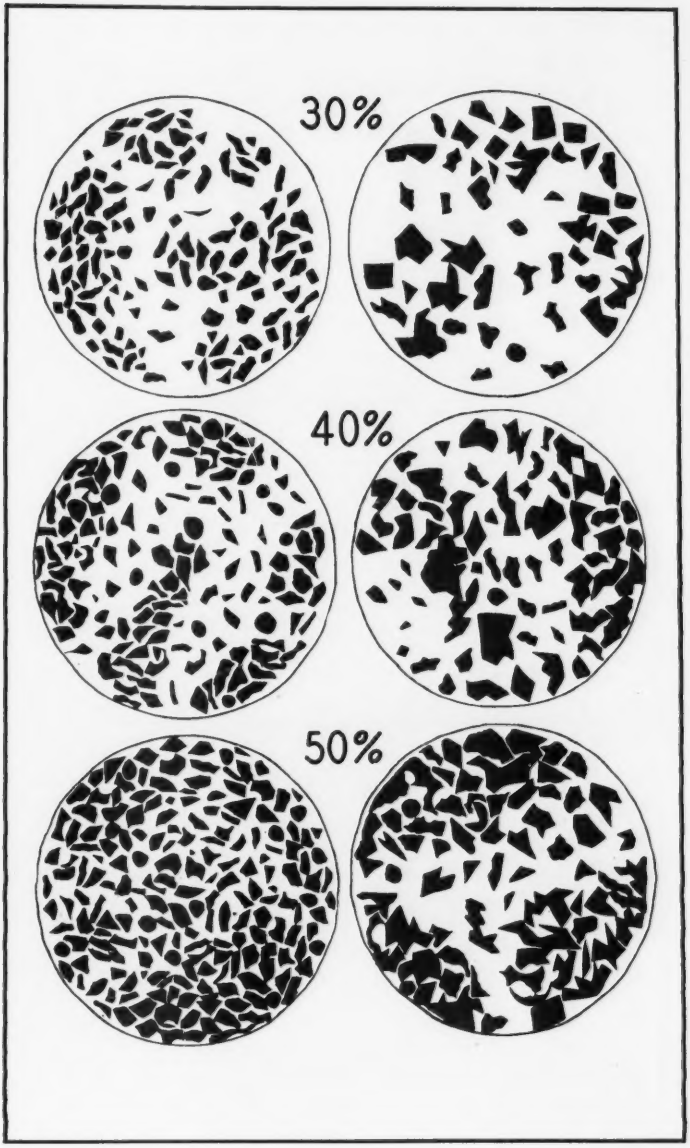

FIG. 4

CONCLUSIONS

Folk (1951) made diagrams similar to Shvetsov's for estimating percentages of minerals etc., in rock sections and also listed additional uses for such diagrams. Grains of more natural shape would probably be more advantageous, but regardless, Shvetsov's and Folk's diagrams are very useful for visual estimation of percentages of grains in rocks.

Griffiths and Rosenfeld (1954) have made an excellent statistical study of operator variations in estimating proportions of minerals, mineral composition, grain size, shape, orientation, density, and porosity. Their study, like Shvetsov's, shows the high variability in estimating various parameters in the study of rock sections.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 25, NO. 3, PP. 235–237
FIGS. 1–2, SEPTEMBER, 1955

HOLLOW SANIDINE GRAINS: STRATIGRAPHIC MARKER FOR THE PIERCE CANYON FORMATION, WEST TEXAS AND SOUTHEASTERN NEW MEXICO

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ABSTRACT

Sand-sized detrital sanidine grains with hollow centers are quite common in the Pierce Canyon formation,[1] a Permian or Triassic redbed of west Texas and southeastern New Mexico. Petrographic examination reveals that this type of hollow feldspar is widespread laterally and that it may be a useful stratigraphic marker.

The Pierce Canyon formation is a red, textually mature siltstone (Folk, 1951) of arkosic composition which overlies the Permian evaporite section in the Delaware Basin of west Texas and southeastern New Mexico. This formation is post-Rustler (upper Permian) and pre-Santa Rosa (upper Triassic) in age and is represented throughout the eastern part of the Delaware Basin by a 350 to 400 foot sequence of brownish-red, thin-bedded, well-sorted siltstone. Interbedded with the siltstone are thin- to very thin-bedded, very fine sandstones. The red color of the sediment is due to hematite-stained illite and to a thin hematite coating on the surface of the detrital silt grains. Gypsum and calcite are the dominant cements. Fresh samples of the Pierce Canyon, obtained from cores, typically contain 50 to 60 percent subangular, plutonic igneous quartz; 15 to 30 percent subangular, brilliantly fresh orthoclase; less than one percent fresh plagioclase; less than one percent well-rounded detrital chert; one to two percent muscovite, biotite, chlorite, and metamorphic rock fragments; one to three percent magnetite-ilmenite, tourmaline, zircon, leucoxene, apatite, and garnet; five to 10 percent hematite and illite; and two to five percent calcite or gypsum cement. The mineral assemblage of this formation does not differ appreciably in any of the hundreds of samples examined from the Delaware Basin. There are local variations in the percentage values of the minerals but these are largely due to difference in texture. The mean grain size of most of the sediment is about 0.04 mm and the extreme range is from clay size to 0.5 mm.

Feldspar grains (predominatly orthoclase) in the Pierce Canyon formation have a mean grain size of 0.04 to 0.05 mm and a maximum of 0.5 mm. They constitute 15 to 30 percent of the sediment and are typically fresh and unaltered with smooth angular to subangular surfaces. The interiors of occasional grains contain a few bubbles (Folk, 1955), usually arranged along cleavage directions, and rarely minute inclusions of rutile. Intermittently dispersed throughout the strata are siltstone lenses less than 10 feet thick which contain detrital sanidine having an entirely different appearance than the more common ortholcase just described. The sanidine is considerably larger than the orthoclase, having a mean grain size of 0.12 to 0.14 mm in all of the samples. Sanidine grains are conspicuous because they contain a large cavity or hole in the midst of an otherwise fresh mineral grain. Their occurrence has been noted in all of the siltstones of the Pierce Canyon and in stratigraphically equivalent beds

[1] The petrology of the Pierce Canyon redbeds is the subject of a Ph.D. dissertation in progress at the Department of Geology, The University of Texas.

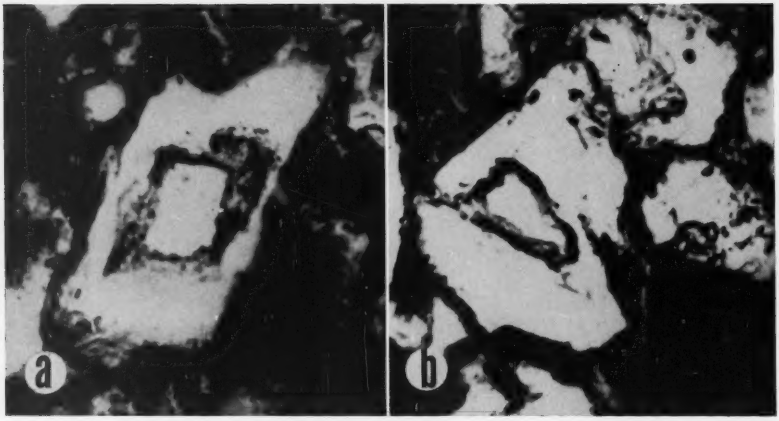

FIG. 1a–b.—Hollow sanidine grains in thin section, Pierce Canyon redbeds, Delaware Basin, Texas and New Mexico (Ordinary light, ×220)

(Dewey Lake formation) farther east. In a few local areas, sanidine constitutes one percent of the total detrital constituents and up to 10 percent of the total feldspar content. In thin section the holes are very conspicuous because they typically occur in the center of the grain as illustrated in figures 1a–b. Most of the holes are empty (except for canada balsam) but a few contain sparry calcite or clear gypsum, the filling corresponding with the cementing constituent in that portion of the rock. Individual grains examined with the binocular microscope typically show an etched external surface and a cleavage controlled lattice or cavity in the center of the grain, which appears white because it is filled with air. Crystals consisting of fragile outer shells with hollow centers have been observed in a few instances where a part of the surface has been broken off, and in some of these the cavity may occupy 80 percent or more of the grain volume. The cavity has a very irregular shape and appears both in binocular and petrographic examination to be controlled more by solution along cleavage planes than by crystal zonation. Most frequently the central cavity follows a pattern similar to the one illustrated in figure 2. In other instances the shape of the cavity is nearly circular. In either case the hole appears to have been dissolved, and cleavage planes with bubbly alteration radiate outward from the hole into the fresh mineral. Elongate grains commonly have two or three holes or cavities. These are usually connected along cleavage planes by a lattice network of minute channels and vacuoles. Grains which do not contain a hole or cavity sometimes exhibit faint crystal zonation which is barely discernable under crossed nicols. Although quite elusive, the difference in optical orientation between the inner and outer zones of the grain is enough to outline a retangular or rhombohedral pattern which is not in optical continuity with the outer shell. Zonation is more evident around the decomposed cavity but only indirectly so because the birefringence blurs as it converges in the immediate area between the hole and the otherwise unaltered sanidine.

The only connecting channels between the hollow center and the fresh exterior surface are the cleavage planes. It is

therefore assumed that fluids penetrated the grain along these cleavage planes and dissolved the less stable central zone. In other instances a fracture or cleaved surface may initially have exposed the unstable interior and solution may have taken place on the surface.

The central holes in the sanidine are believed to have been dissolved after deposition. None of the holes or cavities contain hematite, a few contain calcite or gypsum. It is therefore assumed that the holes were dissolved after deposition, post-hematite precipitation, and precementation. The centrally located holes plus the uniformity of grain size suggests that the sanidine was transported only a short distance. Furthermore, it indicates that the grains have undergone little abrasion and therefore are about the same size as the original crystals in the igneous source.

Sanidine grains with holes or cavities are widely but sparsely distributed throughout the Pierce Canyon formation. Their appearance in this section is so striking that they attract attention. They have not been observed in the underlying redbeds of the Rustler formation or in the overlying Santa Rosa redbeds; hence they may serve a useful purpose as a stratigraphic marker in these otherwise confusing red sediments. In the Delaware Basin region sanidine has been observed in the Pierce Canyon formation as far south as the town of Kent in Culberson County, Texas, as far north as the potash mines in east central Eddy County, New Mexico, and as far east as Colorado City (in Dewey Lake formation) in Mitchell County, Texas.

The origin of the Pierce Canyon sediments has not been determined yet although certain criteria have been established. The preponderance of plutonic igneous quartz, fresh subangular orthoclase, and the accessory minerals previously listed suggest a granitic source area supplemented by very minor amounts of metamorphic and sedimentary materials. Sanidine on the other hand suggests a trachyte or possibly some other eruptive or extrusive rock as a volcanic source. This source may be partially verified because euhedral and embayed quartz, believed to be of volcanic origin, has been observed in a few instances in the thin sections of the Rustler and Pierce Cayon formations. Whether or not the two minerals, sanidine and quartz, originated at the same source is not known. There are two other factors which appear to favor a volcanic source for a part of this sediment: the predominance of biotite over muscovite and the occurrence of apatite as a common heavy mineral. Both minerals, biotite and apatite, might well be related to extrusive rocks. At best, about all that can be said is that volcanic rocks have contributed detritus to this sediment in small quantities, but those small quantities were widely dispersed and their occurrence may serve as a stratigraphic marker for the Pierce Canyon formation.

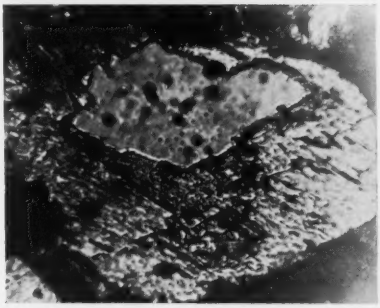

FIG. 2.—Cleavage controlled cavity in hollow sanidine grain (Ordinary light, ×200).

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 25, NO. 3, P. 238
SEPTEMBER, 1955

ANNOUNCEMENT

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The Advisory Committee on Radioactive Mineral Exploration was organized in the summer of 1952. It is composed of 15 members appointed, each for a three-year term, from the A.A.P.G., the S.E.G., and the S.E.P.M.

As stated in the first informational circular distributed by this committee, "It so happens that in many regions where oil and gas are sought, ores of uranium may occur. This is especially true of extensive areas in the Rocky Mountain Province and in the western part of the Central Plains Province. There are many deposits of carnotite (a common uranium ore) in the Plateau Province of southeastern Utah and the adjoining corners of Colorado, New Mexico, and Arizona, and recently deposits of uranium minerals have been found in the Powder River Basin in Wyoming. In these same areas, and in many others where uranium ores *may* occur, exploration for petroleum is being carried on aggressively by numerous companies. Therefore, it is suggested that these companies be encouraged to take steps to combine with their petroleum exploration a moderate effort at exploration for radioactive minerals."

The primary functions of this committee are:

"To invite those who may be inclined or interested, to investigate the feasibility of their carrying on exploration for radioactive minerals along with their petroleum exploratory efforts;

"To give advice as to how and where such exploration for radioactive minerals may be conducted, if desired;

"To advise where information can be obtained concerning the search for radioactive materials;

"To endeavor to keep abreast of what is going on in this search for radioactive materials; and

"To encourage the preparation and presentation of papers on the subject of exploring for useful radioactive minerals.

"This committee will concern itself with over-all encouragement of participation of the oil and gas industry in exploration for radioactive minerals and will deal with the broad aspects and policies of such a relationship. It will not be too much concerned with detailed techniques, problems of instrumentation, methods, and processes for detailed surveys, nor with any of the technical aspects of the program which may better be handled by specialists. In no way will it suggest unified co-operation, or co-operative procedure, between organizations for the conduct of radioactive mineral exploration. Its whole objective is to stimulate an interest in the search for radioactive minerals, occurring within feasible mining depths, when and where petroleum exploration is in progress."

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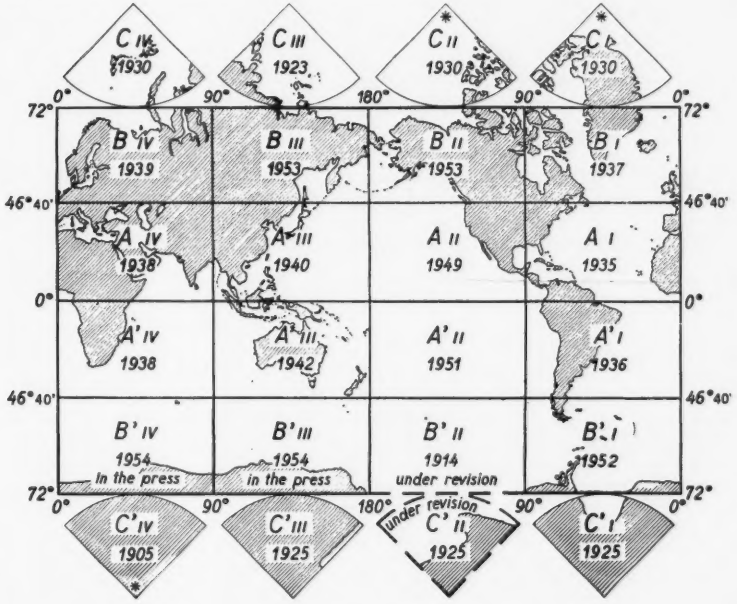

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The Permian Basin Section of the Society of Economic Paleontologists and Mineralogists announces a field trip in the Guadalupe Mountains, October 22 and 23, under the leadership of Phillip B. King. Headquarters for the trip will be Carlsbad, New Mexico. Registration in Carlsbad will be during the afternoon and evening of October 21. Participants will have sufficient time to collect specimens and enter into the discussion of the geology at each stop. Further details can be obtained from Hugh N. Frenzel, secretary of the Permian Basin Section, Box 832, Midland, Texas.

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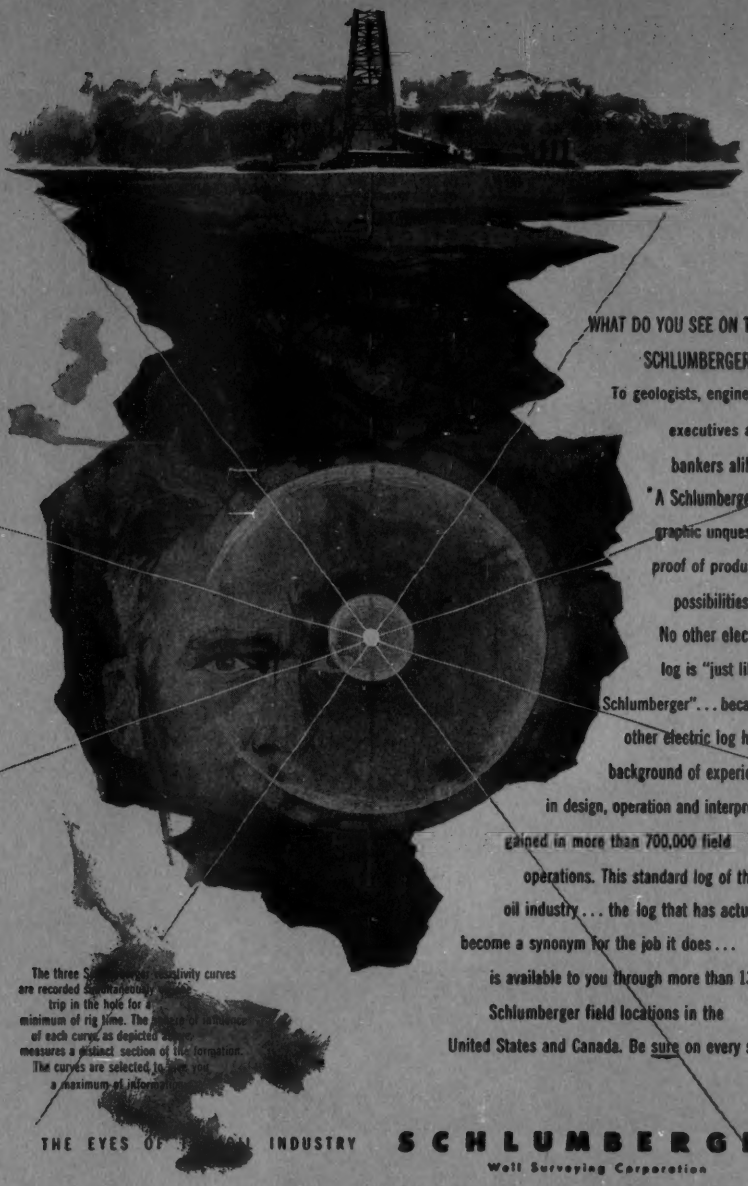

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